UNIVERSITY OF ALBERTA

GLACIAL AND SEA LEVEL HISTORY OF WESTERN FOSHEIM PENINSULA, ELLESMERE ISLAND, ARCTIC CANADA

©

BY

TREVOR BELL

A THESIS
SUBMITTED TO THE FACULTY OF GRADUATE STUDIES AND RESEARCH IN PARTIAL FULFILMENT OF THE REQUIREMENTS FOR THE DEGREE OF DOCTOR OF PHILOSOPHY

DEPARTMENT OF GEOGRAPHY

EDMONTON, ALBERTA
FALL 1992

National Library
of Canada

Acquisitions and
Bibliographic Services Branch

395 Wellington Street
Ottawa, Ontario
K1A 0N4

Bibliothèque nationale
du Canada

Direction des acquisitions et
des services bibliographiques

395, rue Wellington
Ottawa (Ontario)
K1A 0N4

Your file Votre référence

Our file Notre référence

The author has granted an irrevocable non-exclusive licence allowing the National Library of Canada to reproduce, loan, distribute or sell copies of his/her thesis by any means and in any form or format, making this thesis available to interested persons.

L'auteur a accordé une licence irrévocable et non exclusive permettant à la Bibliothèque nationale du Canada de reproduire, prêter, distribuer ou vendre des copies de sa thèse de quelque manière et sous quelque forme que ce soit pour mettre des exemplaires de cette thèse à la disposition des personnes intéressées.

The author retains ownership of the copyright in his/her thesis. Neither the thesis nor substantial extracts from it may be printed or otherwise reproduced without his/her permission.

L'auteur conserve la propriété du droit d'auteur qui protège sa thèse. Ni la thèse ni des extraits substantiels de celle-ci ne doivent être imprimés ou autrement reproduits sans son autorisation.

ISBN 0-315-77417-7

Canadä

UNIVERSITY OF ALBERTA

RELEASE FORM

NAME OF AUTHOR: Trevor Bell

TITLE OF THESIS: Glacial and sea level history, western Fosheim Peninsula, Ellesmere Island, high arctic Canada

DEGREE: Doctor of Philosophy

YEAR DEGREE GRANTED: Fall 1992

Permission is herby granted to THE UNIVERSITY OF ALBERTA LIBRARY to reproduce single copies of this thesis and to lend or sell such copies for private, scholarly or scientific research purposes only.

The author reserves other publication rights, and neither the thesis nor extensive extracts from it may be printed or otherwise reproduced without the author's written permission.

Trevor Bell

12432 52 Ave.,
Edmonton, Alberta
CANADA T6H 0P4

DATE: October 5th, 1992

UNIVERSITY OF ALBERTA

FACULTY OF GRADUATE STUDIES AND RESEARCH

The undersigned certify that they have read, and recommend to the Faculty of Graduate Studies and Research for acceptance, a thesis entitled GLACIAL AND SEA LEVEL HISTORY, WESTERN FOSHEIM PENINSULA, ELLESMERE ISLAND, HIGH ARCTIC CANADA submitted by TREVOR BELL in partial fulfillment of the requirements for the degree of DOCTOR OF PHILOSOPHY.

Dr. John England (Supervisor)

R.B. Rains
Dr. Bruce Rains

John Shaw
Dr. John Shaw

Mark M Fenton
Dr. Mark Fenton

Michael J Retelle
Dr. Mike Retelle

DATE: August 31st, 1992

'TO THE MEMORY OF MY MOTHER'

ABSTRACT

Discontinuous moraines and meltwater channels that either encircle broad uplands or occupy the lower valleys of cirque basins record a restricted ice cover on western Fosheim Peninsula at the last glacial maximum. During deglaciation (8200-9600 BP), meltwater reached the sea, depositing thick sequences of glaciomarine sediments. High sedimentation rates were sustained through the early Holocene warm period, demonstrating the importance of precipitation for the maintenance of high arctic glaciers.

Beyond the last ice limit, there is abundant evidence of older, more extensive glaciations. The oldest recorded advance inundated Fosheim Peninsula and was likely responsible for the deposition of erratics on the highest summits. Fossiliferous till contains shells that have similar amino acid ratios to the Kap København Formation, northern Greenland, dated 2.4-3.2 Ma and to till deposited by the maximum advance of the Greenland Ice Sheet onto northeastern Ellesmere Island. This glaciation may be as old as 2.4 Ma.

During a subsequent glaciation, regional ice sources coalesced on western Fosheim Peninsula. Moraines, meltwater channels and outwash record ice retreat in conjunction with higher relative sea levels between 170 and 185 m asl. The maximum age of this glaciation is estimated to be *ca.* 2 Ma; however, it may be as young as 0.4-0.7 Ma based on paleoclimatic considerations.

Because precipitation is regarded as the critical climate control on modern glacier growth, it has been argued that more extensive glaciations were favoured by much greater precipitation and a marked reduction in the Arctic Ocean sea-ice cover. This relationship appears consistent with the geological evidence for older glaciations on western Fosheim Peninsula.

ACKNOWLEDGEMENTS

Many friends and colleagues have helped me along the PhD road, this last step is a tribute to their encouragement and understanding. Despite suffering an early 'snake-bite' incident, John England has been there every step of the way. John gave me the opportunity to work on Ellesmere Island and, through his dedication and respect for the arctic, has encouraged me to pursue my studies. Thanks John for sharing numerous field trips to Hub mall, tree-planting, grazing in the raspberry patch, open fires under a prairie sky, and your home.

Bruce Rains has been an important part of my academic training at U of A. His support, openness and fairness shine brightly in a sometimes shaded system. Although Don Lemmen left for greener pastures soon after my arrival in Edmonton, our friendship continues to grow. His astute observations, reasoned arguments, editing skills and gentle encouragement gave me a push on some uphill stretches. Doug Hodgson's continual support, advice and friendship have been invaluable through the years. Doug also generously provided unpublished data on Fosheim Peninsula.

Jan Bednarski, Tracy Brennand, Ian Campbell, Bruce Rains and John Shaw are thanked for discussions on various aspects of this research. Richard England, Scott Lamoureux and Gillian Davidge provided dedicated and good-humoured field assistance. Thanks to R. McNeely (GSC) and R. Beukens (IsoTrace) for radiocarbon dating, and Julie Brigham-Grette (University of Massachusetts, Amherst) for amino acid analysis. Bonnie Gallinger made a superhuman effort to draw diagrams under a very tight schedule, with additional help from Lee Martens and Dave Chesterman. Randy Pakan processed the black and white photographs.

Financial assistance for this research was provided by Energy, Mines and Resources research agreement #171, a grant from the Boreal Institute for Northern Studies, a Dissertation Fellowship from the University of Alberta, and NSERC grant #A6680 to John England. Logistical support was provided by the Polar Continental Shelf Project.

My family has been unfailing in their assistance and encouragement through the years; they are an important part of this achievement. To Jean, Barb, Val, Barb D., Lourdes, Bob, Gesche, Bonnie, James, Tracy, Catherine, Wolfgang, Fred, and the Davidges, thanks for the support, the Big Rock and, most of all, the friendship. Finally, thanks to Gillian for enduring 'thesis' life.

TABLE OF CONTENTS

Chapter | Page

1. INTRODUCTION AND GEOLOGICAL BACKGROUND 1
   - 1.1 Introduction 1
   - 1.2 Pre-Quaternary geology 4
   - 1.3 Pre-Quaternary physiography 7
   - 1.4 Tertiary environments 10
     - 1.4.1 Early Tertiary warm period 10
     - 1.4.2 Mid to late Tertiary cooling 11
     - 1.4.3 Early Pleistocene warm period 13
   - 1.5 Earliest regional glaciation 13
   - 1.6 Pliocene transgressions 14
   - 1.7 Pre-Wisconsinan glaciations 17
   - 1.8 Wisconsinan glaciation 18
     - 1.8.1 Ice buildup 21
     - 1.8.2 'Full glacial' sea 21
     - 1.8.3 Deglaciation 23
     - 1.8.4 Postglacial emergence 26
   - 1.9 Holocene environments 27

2. STUDY AREA, RESEARCH OBJECTIVES AND METHODS 28
   - 2.1 Orientation 28
   - 2.2 Geology 31
   - 2.3 Physiography 38
   - 2.4 Climate 42
   - 2.5 Paleoclimate 43
   - 2.6 Local glacial history 44
   - 2.7 Research objectives and rationale 45
   - 2.8 Research methods 47
   - 2.9 Dating former sea levels 48
     - 2.9.1 Elevations of samples and paleoshorelines 48
     - 2.9.2 Sample selection and preparation 49
     - 2.9.3 Radiocarbon dating 50
     - 2.9.4 Amino acid analysis 50

3. SURFICIAL GEOLOGY AND GEOMORPHOLOGY 52
   - 3.1 Weathered bedrock/Residuum 53
     - 3.1.1 Associated landforms 57

Chapter **Page**

3.2 High terrace sediments 59
  3.2.1 Nature and occurrence 59
3.3 Till 63
  3.3.1 Provenance 63
  3.3.2 Distribution 66
  3.3.3 Associated landforms 71
3.4 Glaciofluvial sediments 79
  3.4.1 Distribution 79
3.5 Raised marine sediments 83
  3.5.1 Deltaic, beach, and nearshore sediments 83
  3.5.2 Marine sediment veneer and washed bedrock 86
  3.5.2 Distribution 88
3.6 Fluvial sediments 93
3.7 Minor sediments (not mapped) 93
  3.7.1 Eolian sediment 93
  3.7.2 Organic accumulations 93

4. QUATERNARY HISTORY 95
4.1 Last glaciation 95
  4.1.1 Glacial limit 95
  4.1.2 Marine limit 97
  4.1.3 Deglaciation 98
  4.1.4 Postglacial emergence 99
  4.1.5 Discussion and regional correlation 101
4.2 Previous glaciations 105
  4.2.1 Ice flow patterns and glaciations 106
  4.2.2 Group I glaciation 110
  4.2.3 Group II glaciation 113
  4.2.4 Regional correlation and absolute age estimates 115
  4.2.5 Discussion 119

5. SUMMARY 123

REFERENCES 127

APPENDIX A: AIR PHOTOGRAPHS 144

APPENDIX B: STRATIGRAPHIC SECTIONS 152

APPENDIX C: RADIOCARBON DATES AND AMINO ACID RATIOS 166

LIST OF TABLES

Table		Page
1.1	Circumarctic correlations of Pliocene marine transgressions	15
4.1	Summary group statistics for alle/Ile ratios from the study area	109
4.2	Tentative correlation of late Pliocene/early Pleistocene glaciations and marine episodes in Alaska, Greenland and Ellesmere Island	117
C.1	Radiocarbon dates obtained for this study	167
C.2	Amino acid ratios obtained for this study	170

LIST OF FIGURES

Figure		Page
1.1	Location map of the Canadian Arctic Islands, northern Alaska and Greenland	2
1.2	Location map of the Queen Elizabeth Islands	3
1.3	Geologic provinces of the Arctic Islands	5
1.4	Arctic Tertiary drainage system as proposed by Fortier and Morley (1956)	9
1.5	The Ellesmere-Baffin Glacier Complex	18
1.6	Maximum and minimum extent of the last glaciation, Queen Elizabeth Islands	19
1.7	General limit of the last glaciation on northern and western Ellesmere Island	22
1.8	Isobases drawn on the 10 000 BP shoreline on northeastern Ellesmere Island	23
1.9	Limit of the last glacial maximum in Greely and Cañon fiords.	24
1.10	Provisional emergence curve for inner Greely fiord	26
2.1	Eastern Queen Elizabeth Islands showing location of study area	29
2.2	Official and unofficial placenames of western Fosheim Peninsula	30
2.3	Stone cache and tent rings from a seasonal hunting camp built by Thule Inuit	32
2.4	Eureka weather station on the northern shore of Slidre Fiord	32
2.5	Structural cross-section AA' through Fosheim Peninsula	33
2.6	Simplified geology map of western Fosheim Peninsula	35
2.7	Strike-aligned scarp-ridges on the western flank of Airfield ridge	36
2.8	Iceberg Bay Formation displaying sandstone-coal cycles	36
2.9	Erosional remnant of a Neogene high terrace	37
2.10	Thrust sheets in the Sawtooth Range	37
2.11	Snow-covered, Landsat image of Fosheim Peninsula	39
2.12	Ridge crest of the Sawtooth Range	40
2.13	View southwest across a high terrace remnant in Wolf Valley	40
2.14	View looking NNE across Slidre Fiord towards Black Top Ridge	41
2.15	View across till-blanketed Iceberg Point lowlands towards Mount Lockwood	41

3.1 Surficial geology and geomorphology, western Fosheim Peninsula *rear pocket*
3.2 Resistant sandstone ridge capping recessive shale, south of Slidre Fiord 54
3.3 Incipient stone pavement in residuum near Hare Cape ridge 54
3.4 Cavernous weathering (tafoni) in sandstone near Hare Cape ridge 56
3.5 Sandstone tor in the central lowlands of western Fosheim Peninsula 58
3.6 Nodular weathering on sandstone 58
3.7 Isolated remnant of high terrace sediment 60
3.8 Edge of till-covered plateau exposes high terraces sediments 60
3.9 Meltwater channel eroded in high terrace sediments 61
3.10 Location of shell sites, stratigraphic sections and marine limit *rear pocket*
3.11 Examples of unaltered wood associated with or recycled from high terraces 61
3.12 Till blanket obscuring underlying bedrock structure on the central lowlands 64
3.13 Till veneer on Weather station lowland 64
3.14 Discontinuous till veneer on the eastern slope of Troelsen ridge 65
3.15 Typical appearance of quartzite erratics in the study area 65
3.16 Distribution of granite erratics along northern Eureka Sound 67
3.17 Large granite boulder near Blue Man Cape 68
3.18 Slumped exposure of granite-rich till on Iceberg Point lowland 68
3.19 Oblique view of washed moraine or grounding line near Iceberg Point 72
3.20 View looking east through northernmost valley draining Troelsen ridge 72
3.21 View looking west across till and an end moraine near Black Top Ridge 74
3.22 View looking upvalley at moraines on Hare Cape ridge 74
3.23 Oblique views of Romulus moraine looking southwest 75
3.24 Boulders eroded from Romulus moraine armour Slidre River channel 77
3.25 Photograph of lateral moraine oriented parallel to Eureka Sound 77
3.26 View towards the Sawtooth Range where it intersects Eureka Sound 78
3.27 Coarse gravel lag on the floor of meltwater channels eroded in till 78
3.28 View across discontinuous kame moraine on Weather station lowland 80
3.29 Dissected surface of delta at the northern end of Black Top Ridge 80
3.30 Dissected outwash above Big Slide creek 82
3.31 Paleochannels preserved on the outwash surface in Big Slide creek valley 82
3.32 Delta foresets inclined at 17-22° in Eureka Sound 84
3.33 Sea-ice pushed ridge composed of sea ice and sediment at Iceberg Point 84
3.34 Lithofacies **Ia** consisting of horizontally and thinly bedded mud 85
3.35 Lithofacies **Ib** consisting of horizontal, graded beds of sand and silt 85
3.36 Lithofacies **II** consisting of graded beds of fine sand, silt and clay 87
3.37 Massive ice body underlying lithofacies **Ia** marine sediment 87
3.38 Oblique view of marine limit near Remus Creek 90
3.39 Outcrop of fine-grained marine sediment on Greely Fiord coast 90
3.40 View looking west across planed outwash surface with beach berms 91
3.41 View across the flat surface of marine sediments in Slidre River Valley 91
3.42 A thick section of marine sediments in Slidre River valley 92
3.43 The lower limit of outwash marks marine limit in Fox river valley 92

4.1 Paleogeography of western Fosheim Peninsula during the last glaciation 96
4.2 Minimum emergence curve for northern Fosheim Peninsula 100
4.3 Plot of free vs. total alle/Ile ratios from shells in the study area 107
4.4 Frequency distributions of free and total alle /Ile ratios 108
4.5 Paleogeography of western Fosheim Peninsula during Group I glaciation 111
4.6 Paleogeography of western Fosheim Peninsula during Group II glaciation 114

A.1 Vertical air photograph of northern Black Top Ridge and adjacent lowlands 145
A.2 Oblique air photograph looking NW towards Nansen Sound 146
A.3 Oblique air photograph looking SE towards Cañon Fiord 147
A.4 Vertical air photograph of the coastal lowlands near Greely Fiord 148
A.5 Vertical air photograph of Iceberg Point lowland 149
A.6 Oblique air photograph looking SSE towards central Eureka Sound 150
A.7 Oblique air photograph looking SE towards Slidre River Valley 151

B.1 Section 1 near Cape Lockwood 153
B.2 Close-up of section 1 153
B.3 Section 2 near Snow Goose hill 155
B.4 Close-up of section 2 155
B.5 Section 3 near Iceberg Point 156
B.6 View of till-sand contact in section 3 156
B.7 Schematic diagram of section 4 in tributary valley of Romulus river 158
B.7 General view of section 4 in tributary valley of Romulus river 159
B.8 View of the lowermost sand and the overlying till in section 4 159
B.9 View of interbedded debris flows and channel/tunnel sand in section 4 160
B.10 View of the central part of section 4 showing interbedded debris flows 160
B.11 View of the upper till that caps section 4 162
B.12 View of the penetratively deformed zone below the upper till in section 4 162
B.13 Section 5 exposed in a retrogressive thaw slump near Hot Weather Creek 164
B.14 View of section 5 showing sand rafts in till matrix 164
B.15 Close-up of large sand raft in section 5 165
B.16 Close-up of stacked sand rafts in section 5 165

CHAPTER ONE

Introduction and geological background

1.1 INTRODUCTION

This introductory chapter summarizes the recent geologic and climatic evolution of the high arctic which is fundamental to our understanding of the glacial and sea level history of western Fosheim Peninsula. Northern and western Ellesmere Island provide one of the most detailed environmental records bordering the Arctic Ocean basin (Fig. 1.1). During the past 15 years, researchers from the University of Alberta have documented the late Quaternary glacial and sea level history of this region. Collectively, they have mapped a discontinuous ice cover for the last glaciation, termed the Franklin Ice Complex (England 1976*a*, 1983, 1990; Bednarski 1986; Lemmen 1989; Evans 1990; Sloan 1990). Their reconstruction challenges the model of a more extensive regional ice cover, the Innuitian Ice Sheet, proposed by Blake (1970) and accepted by others (Hughes *et al.* 1977; Mayewski *et al.* 1981; Tushingham 1991). Ironically, neither of these opposing models has been tested at the centre of the proposed ice sheet. Western Fosheim Peninsula, bordering Eureka Sound, provides this opportunity (Fig. 1.2).

Details on pre-late Quaternary glaciation and sea level history remain limited despite the extensive fieldwork of the last 15 years. However, earlier reports of high-elevation shells, far-travelled erratics and meltwater channels on local ridge summits suggest that western Fosheim Peninsula may preserve a record of older glaciations (Sim 1961; Hodgson 1985). Other enigmatic features of the peninsula which must be reconciled with any glacial reconstruction include: *in situ* surface tree stumps from a Tertiary temperate forest; remnants of late Tertiary fluvial planation; and the highest postglacial emergence on Ellesmere Island. Although the magnitude of this emergence is consistent with a regional ice cover, its pattern is unlike any drawn for glaciated areas of arctic Canada (England 1992).

In this thesis, I present the glacial geology, geomorphology, stratigraphy and relative sea level record of western Fosheim Peninsula and provide a chronology of Quaternary glaciations and marine transgressions based on relative position,

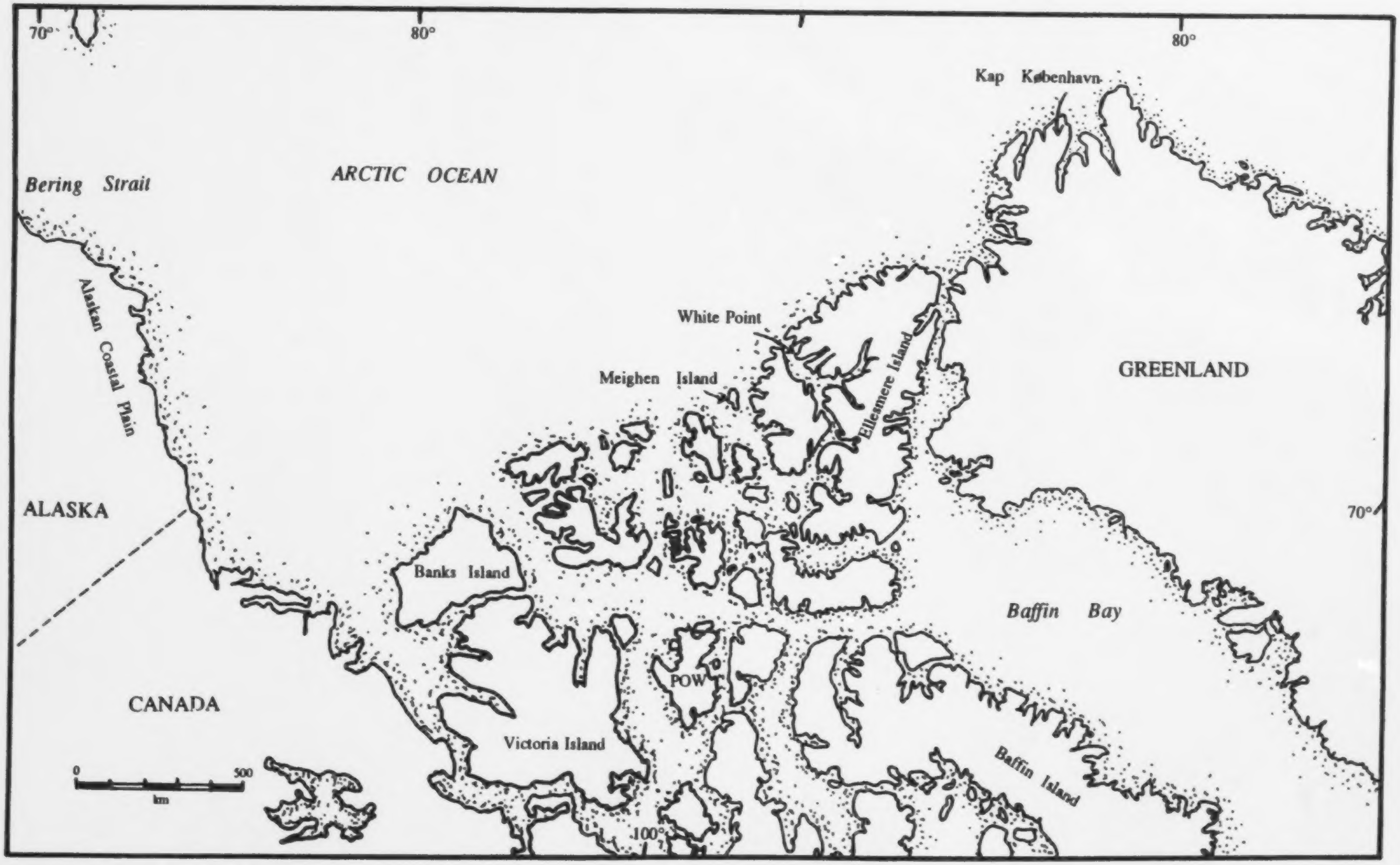

70°
80°
80°
70°
100°
Kap København
Bering Strait
ARCTIC OCEAN
Alaskan Coastal Plain
ALASKA
CANADA
White Point
Meighen Island
Ellesmere Island
GREENLAND
Banks Island
Victoria Island
POW
Baffin Bay
Baffin Island
0
500
km

Figure 1.1: Location map of the Canadian Arctic Islands, northern Alaska and northern Greenland.

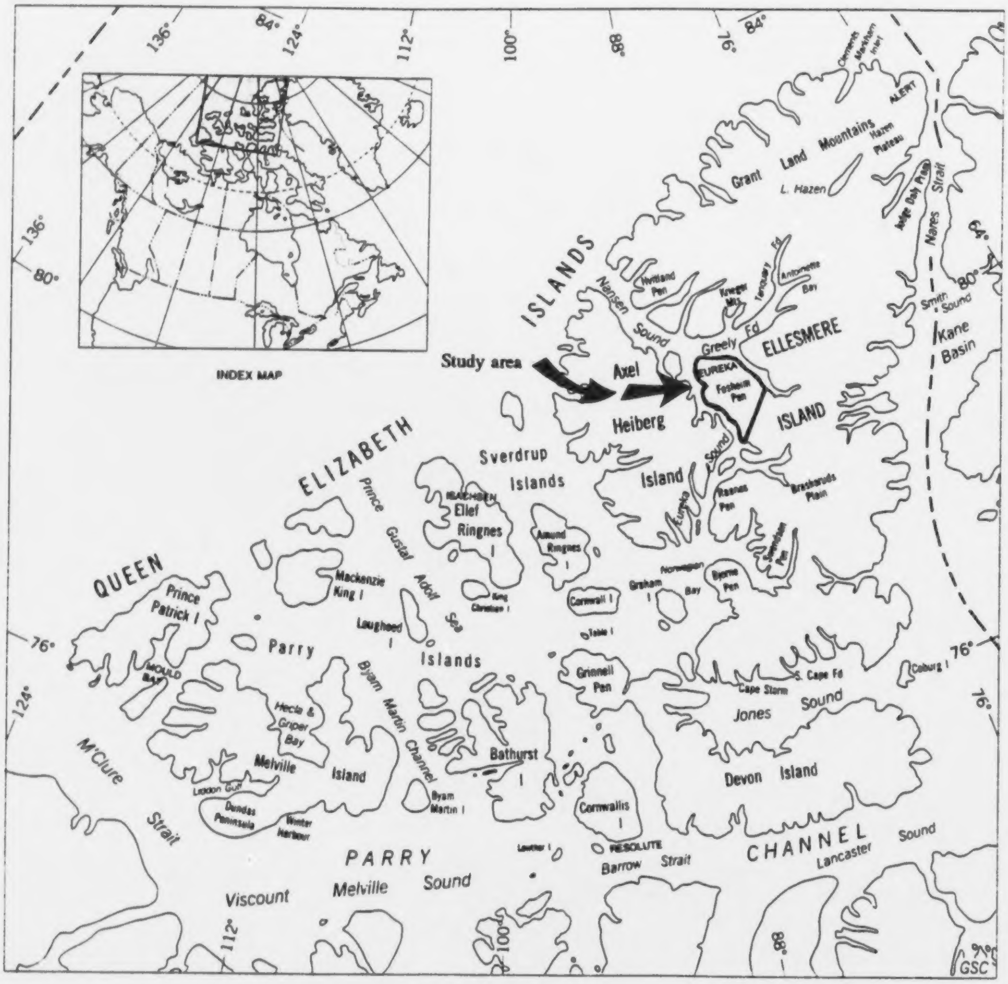

INDEX MAP
Study area
QUEEN ELIZABETH ISLANDS
Axel Heiberg Island
ELLESMERE ISLAND
EUREKA
Fosheim Pen
Grant Land Mountains
Hazen Plateau
L. Hazen
ALERT
Nansen Sound
Greely Fd
Antoinette Bay
Nares Strait
Smith Sound
Kane Basin
Sverdrup Islands
ISACHSEN
Ellef Ringnes I
Amund Ringnes I
Prince Gustaf Adolf Sea
Mackenzie King I
Prince Patrick I
MOULD BAY
Parry Islands
Byam Martin Channel
Hecla & Griper Bay
Melville Island
Dundas Peninsula
Winter Harbour
Byam Martin I
Bathurst I
Cornwallis I
RESOLUTE
Cornwall I
Graham I
Norwegian Bay
Bjorne Pen
Grinnell Pen
Jones Sound
Cape Storm
S. Cape Fd
Coburg I
Devon Island
M'Clure Strait
Viscount Melville Sound
PARRY CHANNEL
Barrow Strait
Lancaster Sound
Lowther I
Table I
Eureka Sound
Raanes Pen
Svendsen Pen
Bruce Pen
Lougheed I
Liddon Gulf
King Christian I
136°
124°
112°
100°
88°
76°
84°
80°
76°
64°
GSC

Figure 1.2: Location map of the Queen Elizabeth Islands (adapted after Hodgson 1989).

radiocarbon dating and amino acid analysis of organic remains. In addition, I attempt a circumarctic correlation of glacial and marine events using the Fosheim chronology and recently published data from Alaska and northern Ellesmere Island (Brigham-Grette and Carter 1992; Lemmen and England 1992). The implications of this research are discussed with reference to the geologic and paleoclimatic evolution of the high arctic.

In 1988, the Geological Survey of Canada (GSC) established the Global Change high arctic observatory on western Fosheim Peninsula. Although this research is not part of the global change project, it contributes to our understanding of high latitude paleoenvironmental change in several ways: (1) it shows the extent over which the high arctic environment has changed both in the last few thousand years and in the last few million years; (2) it provides a record with which to test global models now being constructed by atmospheric and oceanographic scientists (*e.g.* the Cooperative Holocene Mapping Project [COHMAP] 1988); and (3) it demonstrates that the high arctic can be extremely sensitive to environmental thresholds which propagate their effects through the Earth system. (*e.g.* abrupt changes in climate and the glacial system).

Chapter 2 introduces the study area and outlines the research objectives and methods. Chapter 3 constitutes the main data source on which the thesis is based. It describes the surficial geology and geomorphology of the study area and is supplemented by other data in appendices. The Quaternary history chapter (4) provides a synthesis of the glacial and sea level history of western Fosheim Peninsula and develops a regional context for the correlation of circumarctic events. Finally, the implications of the research are presented and discussed. Chapter 5 presents the conclusions.

1.2 PRE-QUATERNARY GEOLOGY

The bedrock of western Fosheim Peninsula is dominated by sedimentary rocks of the Sverdrup Basin which characterized the late Early Carboniferous to Early Tertiary history of the Canadian Arctic Islands. This basin was one of a series that developed on the Precambrian basement; from oldest to youngest they include: an unnamed Proterozoic basin; the Franklinian Mobile Belt and Arctic Platform; the Sverdrup Basin; and the Arctic continental terrace wedge (Fig. 1.3).

The Precambrian crystalline basement crops out as the **Canadian Shield** on southeastern Ellesmere Island, eastern Devon Island, and Baffin Island (Fig. 1.3). It extends under and, in places, is exposed through the adjacent Phanerozoic sedimentary cover (*e.g.* Boothia Uplift); however, its northern limit in the subsurface has yet to be determined (Trettin 1989). Preliminary structural sections of western Fosheim Peninsula suggest that the depth to basement is about 15-18 km, which

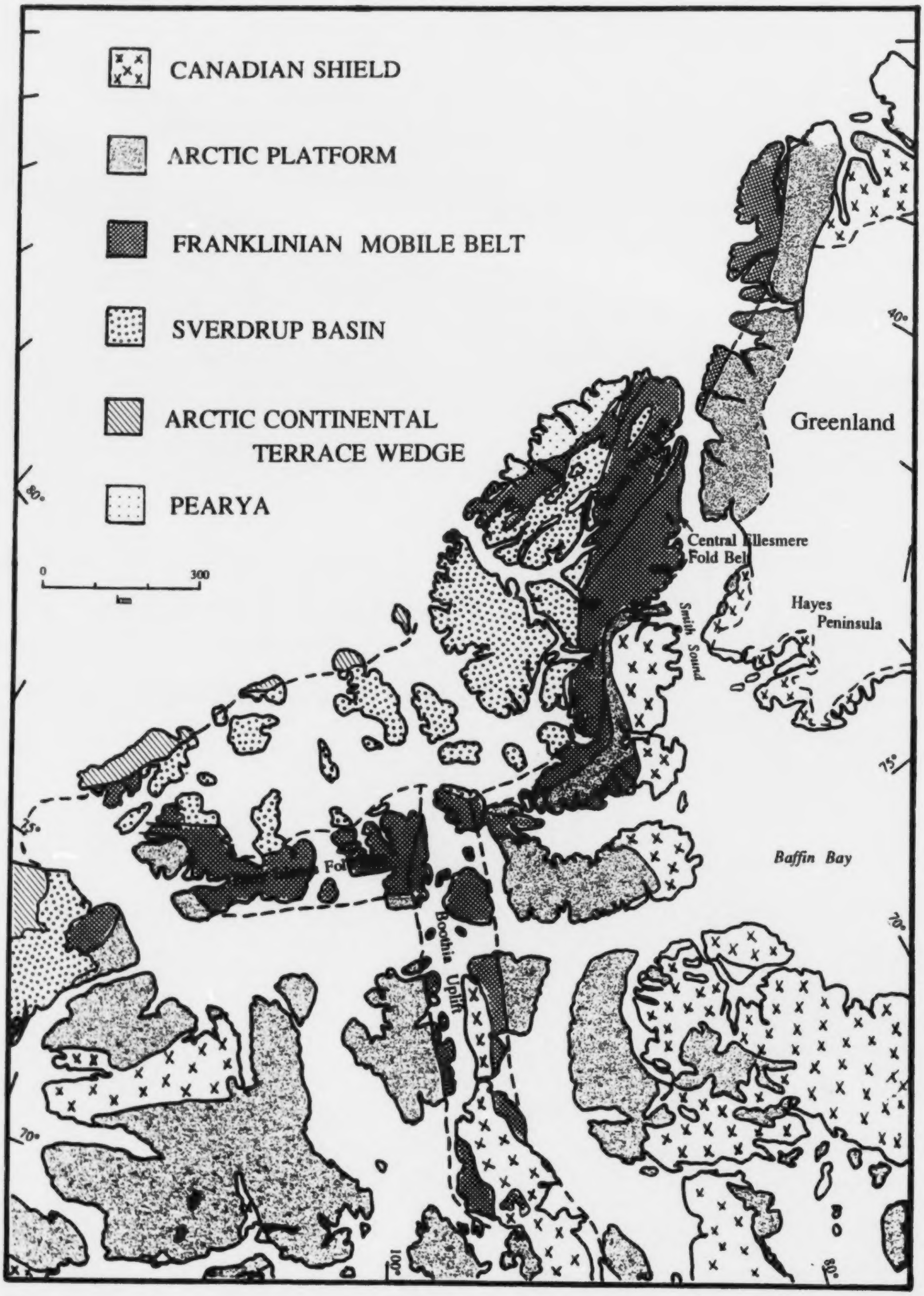

CANADIAN SHIELD
ARCTIC PLATFORM
FRANKLINIAN MOBILE BELT
SVERDRUP BASIN
ARCTIC CONTINENTAL TERRACE WEDGE
PEARYA
0
300
km
Greenland
Central Ellesmere Fold Belt
Hayes Peninsula
Smith Sound
Baffin Bay
Boothia Uplift
40°
80°
75°
70°
100°
80°

Figure 1.3: Geologic provinces of the Arctic Islands.

roughly corresponds to estimates based on geophysical parameters for central Queen Elizabeth Islands (Sobczak *et al.* 1986; Ricketts 1987*a*).

The largest sedimentary basin is composed of the **Franklinian Mobile Belt** and its associated shelf, the **Arctic Platform** (Fig. 1.3). Here, Cambrian to Devonian strata were deposited in shelf and deep water basins along the northwest margin of the former craton. The Franklinian Shelf and Arctic Platform are stratigraphically continuous; the boundary between them coincides with the southern limit of folding during the Ellesmerian Orogeny. On northern Ellesmere Island, there lies an extensive sedimentary-volcanic terrain of Middle Proterozoic to Late Silurian age. Trettin (1987) proposed that this terrain, called **Pearya,** was a fragment of the Caledonian mobile belt which was rafted by strike-slip motion and accreted to the Franklinian mobile belt during the Late Silurian (Fig. 1.3).

Widespread uplift and folding during the mid-Palaeozoic Ellesmerian Orogeny terminated the Franklinian Mobile belt and produced a regional angular unconformity throughout the Queen Elizabeth Islands.

The **Sverdrup Basin** developed as a rift on the uplifted and eroded Franklinian Mobile Belt in the Early Carboniferous, eventually extending from Ellesmere Island to southwestern Prince Patrick Island (Fig. 1.3). Overall, the basin-fill consists of clastic sediments (up to 8 km thick on Fosheim Peninsula) derived from sources to the south and east. Continuous subsidence and sedimentation occurred over 260 Ma until the Late Cretaceous when the first phase of the Eurekan Orogeny fragmented and folded the basin. The Eurekan Orogeny and the Eurekan Rifting episode are two related and complementary structural events that compose the Eurekan Deformation (Kerr 1981). Eurekan Rifting originated in the southeast arctic islands and advanced northwestward by propagation of extension faults as Greenland and the rest of North America moved apart. The Eurekan Orogeny was a compressional deformation located northwest of the area of rifting (Thorsteinsson and Tozer 1970). Balkwill (1978) identified three phases of deformation: **(1)** uplift and erosion over intrabasinal arches; **(2)** thrust faulting on Ellesmere and Axel Heiberg islands; and **(3)** localized normal faulting. Phases 1 and 2 have been widely recognized in the Queen Elizabeth Islands but Phase 3 has not been widely recognized although it has been supported on geomorphological grounds (Bird 1967; England 1987; Dyke *et al.* 1991). Kerr (1981), however, described the final phase of the Eurekan Deformation (Miocene/Pliocene) as a period when extension was prevalent throughout the Canadian Arctic Islands and Eurekan Rifting superseded the Eurekan Orogeny.

In the Sverdrup Basin, the Eureka Sound Group was deposited in structural depressions during regional uplift and erosion. The stratigraphy is controversial and difficult to resolve because of complex facies relations (Trettin 1989), however, it is generally recognized that deposition took place in predominantly fluvial and deltaic environments, under unstable tectonic conditions (Miall 1986, 1988; Ricketts 1988, 1991; Ricketts and McIntyre 1986). The climactic phase of the Eurekan Orogeny (mid Eocene to early Oligocene) involved compressional deformation which folded the Eureka Sound Group in eastern parts of the Sverdrup Basin.

Deformed Sverdrup Basin strata are largely overlain by Miocene and Pliocene deposits that form part of the **Arctic continental terrace wedge**. Uplift and erosion of the arctic landmass were isostatic responses to Eurekan crustal thickening and caused the main locus of sedimentation to shift to the continental margin (Fig. 1.3). The absence of Oligocene deposits in the regional sedimentary record suggests that this was a period of widespread uplift and erosion and therefore, they may be preserved in the thick offshore record of the continental terrace wedge (>3000 m thick off Meighen Island; Asudeh *et al.* 1989). Onshore sediments consist typically of unlithified, flat-lying, quartz-rich, fossiliferous sand and gravel preserved in a belt 1200 km long by 50 km wide, from Meighen Island to southern Banks Island (Tozer 1956; Tozer and Thorsteinsson 1964). Although the name Beaufort Formation has been loosely applied to Tertiary and Quaternary sediments of the central and western Arctic (Tozer 1956), recent investigations have determined that the title should be restricted to its type area on Prince Patrick Island and demonstrated equivalents (Fyles 1990). Late Tertiary deposits are present on a number of islands east of the Arctic continental terrace wedge. In some cases, they are called Beaufort Formation, but Fyles (1990) strongly recommends alternative nomenclature. For instance, Fyles (1989) described Neogene deposits on Fosheim Peninsula as *high terrace sediments*.

At its type section, the Beaufort Formation is a single lithostratigraphic unit, a few tens of metres thick, consisting of sandy, braided alluvium (Fyles 1990). Organic beds in the sand likely originated during the Pliocene when the area supported a coniferous forest (Fyles 1990; Matthews and Ovenden 1990; Devaney 1991). Unconsolidated sediments on islands north and south of Prince Patrick Island appear to be stratigraphically equivalent to the "type" Beaufort Formation whereas others differ both stratigraphically and paleontologically and may be early to middle Miocene or Pleistocene (Fyles 1990).

1.3 PRE-QUATERNARY PHYSIOGRAPHY

Bird (1967) described the main physiographic elements of the Canadian Arctic as discontinuous erosion surfaces at various elevations on the islands and mainland. These surfaces are interrupted by marine channels that extend to the continental shelf from the interior of the archipelago (Fig. 1.2). The channels are commonly >100 km wide and 500 m deep and maintain a near constant width along their length. They postdate the formation of the erosion surfaces and are younger than the Pliocene Beaufort Formation.

Bird (1967) recognized that the erosion surfaces were formed on a contiguous landmass which was subsequently rifted to produce the intervening channels. Until recently, however, no attempt was made to integrate these surfaces with the sedimentary record. Dyke *et al.* (in press) have identified five pre-Quaternary, rock-cut, planation surfaces on Prince of Wales Island, ranging in elevation from the tilted Barrow Surface (200-400 m above sea level [asl]) to the Prince of Wales Lowland (50 m asl). These surfaces are separated in places by large valleys, 7-13 km wide, 20-30 m deep, with meanders of tens of kilometres wavelength. The youngest valleys appear to be coeval with the Prince of Wales Lowland which Dyke *et al.* (in press) correlate with deposition of the Eureka Sound Group, the youngest sediments of the Sverdrup Basin. They postulated that the wide, low-gradient rivers which flowed across Prince of Wales Lowland must have been comparable in size to the largest rivers of North America today and likely carried only suspended sediments and fine bed load. Consequently, these rivers could not have deposited Beaufort sand and gravel. Because lower erosion surfaces and paleochannels were not observed on Prince of Wales Island, Dyke (pers. comm. 1991) proposes that the Beaufort Formation had local sources in highlands adjacent to the Arctic coastal plain (*e.g.* on Banks, Victoria, Melville and Axel Heiberg islands).

The origin of the inter-island channels is central to any discussion of the physiography of the Canadian Arctic (England 1987). Two hypotheses are commonly presented to explain their evolution: one invokes fluvial dissection during the Tertiary followed by glacial modification in the Pleistocene (Fortier and Morley 1956; Pelletier 1966; Ricketts 1987*b*; Trettin 1989); the other relies on Tertiary tectonism (Bird 1967; Kerr 1980; England 1987). As stated previously, the channels are younger than the Pliocene Beaufort Formation but their relationship to Pleistocene glaciation is unresolved (*cf.* Pelletier 1966; MacLean *et al.* 1987), even though they are frequently used to infer glacial style (Blake 1970; Sugden 1977; CLIMAP 1976; Denton and Hughes 1981).

Fortier and Morley (1956) proposed that the pattern of marine channels was a product of a preceding dendritic drainage, developed on a contiguous landmass since the Tertiary (Fig. 1.4). They defined an Arctic and an Atlantic watershed by postulating a main divide approximately bisecting the Archipelago. Dyke *et al.* (in press) outline the main problems inherent in the fluvial hypothesis:

- Drainage patterns other than dendritic can be drawn through the channels;
- The morphology and scale of the channels are unlike any valley of known fluvial origin elsewhere in the world;
- The floors of some channels extend >500 m below sea level and therefore, during a post-Beaufort interval, sea level must have fallen by >500 m to allow valley incision. Furthermore, a transgression of equivalent magnitude would be necessary to bring the sea up to its present level. There is no independent evidence for such sea level adjustments since the Pliocene.

- According to the fluvial hypothesis large deltaic accumulations roughly equivalent to channel volume should be present at the mouth of each channel, yet such deposits have not been observed.

Some advocates of the fluvial hypothesis invoke glacial erosion to account for the morphology and size of marine channels (*cf.* Ricketts 1987*b*; Trettin 1989). Again, there is no independent evidence to support such glacial modification and voluminous deposits of glacial origin have not been recorded beyond the channels (England 1987). The geomorphological analyses of Pelletier (1966) and White (1972, 1988) are frequently cited as evidence for glacial erosion but their arguments have been accused of "gigantism" (Dyke *et al.* in press). For example, Pelletier describes Prince Gustaf Adolph Sea as a 160 km wide by 400 m deep U-shaped valley. Such an appearance relies on diagrammatic vertical exaggeration. Likewise he refers to drowned cirques with sidewalls 30 km apart. Normal U-shaped valleys and cirques are orders of magnitude smaller.

REMOVED DUE TO COPYRIGHT RESTRICTIONS

Figure 1.4: Arctic Tertiary drainage system as proposed by Fortier and Morley (1956) and represented by England (1987). According to their model, present fiords and marine channels represent submerged dendritic drainage.

Bornhold *et al.* (1976) used detailed bathymetric maps to define five distinct drainage basins on the floor of Barrow Strait which apparently substantiated Fortier and Morley's fluvial hypothesis. However, their data can equally support a tectonic hypothesis of channel origin and, at least locally, the data dismiss the possibility of deep glacial erosion of a pre-Quaternary drainage.

Although Bird (1967) outlined a general hypothesis on the origin of the inter-island channels by rifting, it was Kerr (1981) who presented precise details on the rifting episode. He proposed that the present physiography resulted largely from geological structures that were produced by the Eurekan Deformation and subsequently modified by erosion. One of the final events of the Eurekan Deformation was the breaking-through of faults in Nares Strait and Parry Channel to isolate the completely fault-bounded Queen Elizabeth Islands subplate. Other low-order subplates are separated by faults within the archipelago and these faults, according to Kerr (1981), determine the shapes of some present islands and channels. For example, Somerset Island is a horst, almost completely surrounded by normal faults (Fig. 1.4; Kerr and deVries 1977). England (1987) advanced the argument for the tectonic origin of the arctic islands by precluding glacial erosion as a causal mechanism for the high arctic channels, such as Greely Fiord/Nansen Sound, which also do not conform to the late Tertiary drainage. He proposed that the ongoing tectonic evolution of the arctic islands may have constrained glaciation through time.

1.4 TERTIARY ENVIRONMENTS

1.4.1 Early Tertiary warm period

The occurrence of a "temperate" flora in the high arctic during the early Tertiary has long being recognized although its paleoclimatic significance remains unresolved (Wolfe 1978, 1980; Axelrod 1984; Barron 1985; Basinger 1986, 1991; McMillan 1986). A predominantly deciduous, Arcto-Tertiary vegetation (*ca.* 45 Ma; Basinger 1991) has been described from Axel Heiberg and Ellesmere islands (McMillan 1963; Francis 1988; Basinger 1991). The Axel Heiberg forest grew on a swampy floodplain between large rivers (Ricketts 1991). Alluvium protected the leaf and wood from decay, preserving spectacular, "mummified" forests in rooted position. These stumps are the remains of trees that were ≥1 m in diameter and 20-30 m high (Francis 1988, 1991). Tree species included dawn redwood, swamp cypress, pine, fir, larch, walnut, birch and oak (Basinger 1986, 1991).

According to Irving and Wynne (1991), the paleolatitude of this Eocene forest was only 2° farther south of its present location. As a result, the flora tolerated a polar regime of winter darkness, intervening months of diffuse light and summer months of continuous sunlight. Nevertheless, the climate was clearly suitable for forest growth. As is the case today, the annual amount of sunlight in polar regions is equivalent to that received at lower latitudes because the sun shines for such long hours in summer (Creber and Chaloner 1985). The trees apparently were able to grow continuously during the warm summer season but because of their deciduous nature, they could shed foliage and become dormant during the dark winter months (Axelrod 1984).

Floristics and wood anatomy of the fossil forest indicate that the polar Eocene climate was mild and equable with winter temperatures seldom below freezing (Basinger 1986, 1991). The occurrence of a diverse Eocene vertebrate fauna on Ellesmere Island (*e.g.* flying lemurs, salamanders, lizards, snakes, alligators and turtles) supports the concept of mild polar winters (Dawson *et al.* 1976; West *et al.* 1977; McKenna 1980). Abundant rainfall throughout the year was also a feature of this climate as inferred from vegetation studies, coal deposits (Ricketts and Embry 1984), and analysis of local paleosols (Tarnocai and Smith 1991). The Arctic Ocean, which was largely land-locked and warm (*ca.* 15°C according to Shackleton and Kennett (1975)) at this time, would have moderated winter temperatures (Basinger 1991).

1.4.2 Mid to Late Tertiary cooling

The onset of Tertiary cooling in the Late Eocene had a profound effect on the northern forests. The cause of this climate change was likely a combination of the effects of continental drift and tectonic evolution (Broecker *et al.* 1985; Ruddiman and Raymo 1988; Ruddiman and Kutzbach 1989; Repenning 1990), lowering of high atmospheric CO_2 levels, and orbital forcing (Imbrie and Imbrie 1980; Berger 1984). As Earth's supercontinents were tectonically rearranged, the strong meridional heat transport of the early Tertiary was replaced by a more zonal paleocirculation with less heat transported poleward (Berggren 1982). This was a critical factor in high latitudes where, during winter darkness, net radiation loss was greatest and temperatures were dependent on the rate of heat transfer to the poles.

Decreasing winter temperatures and progressive shortening of the growing season favoured the growth of evergreen conifers. By the Miocene, a mixed evergreen coniferous/broad-leaved deciduous forest had been established (Basinger 1991). Remains of this flora have been found on Banks Island (Hills and Ogilvie 1970; Hills and Matthews 1974; Matthews and Ovenden 1990). Meanwhile, the Pliocene Beaufort Formation on Prince Patrick Island contains plant and insect fossils indicative of coniferous forest whereas the Meighen Island Beaufort flora represent forest tundra (Hills 1975; Matthews 1976; Matthews and Ovenden 1990). Hills (1975) suggested that these floristic distinctions were not evolutionary in nature but, instead, represent a latitudinal gradient with the more northerly sites containing the most impoverished flora. Alternatively, such distinctions could arise from differences in the continentality of sites (Matthews and Ovenden 1990), especially considering that the Beaufort Formation was deposited on a formerly contiguous landmass which was disrupted by block-faulting, perhaps helping to displace the boreal forest southward (England 1987). However, there is evidence to support a general cooling trend in the arctic during the late Tertiary:

(1) Ice-rafted debris in Norwegian Sea (≥2.8 Ma) and Arctic Ocean (4.5-2.5 Ma) sediments and glacial erratics in Alaskan transgressive sediments (2.7-2.48 Ma) indicate at least local glaciation in Scandinavia and North America, respectively (Herman and Hopkins 1980; Jansen and Sjøholm 1991; Brigham-Grette and Carter 1992).
(2) The earliest unequivocal record of permafrost consisted of ice-wedge casts in sediments associated with tundra vegetation in northeastern Siberia *ca.* 2.5 Ma (Sher *et al.* 1979). Meanwhile, Brigham-Grette *et al.* (1987) discussed the likely occurrence of permafrost on Meighen Island shortly after the Beaufort marine transgression (*ca.* 3 Ma, Fyles *et al.* 1991) on the basis of calculated diagenetic temperatures neede to produce the amino acid ratios in Beaufort mollusks.
(3) Although the sediment record of the central Arctic Ocean has led to the development of diverse interpretations of sea ice cover, there is general agreement that the Arctic Ocean suffered a sharp decrease in temperature during the Pliocene (Herman and Hopkins 1980; Clark 1982).

Initiation and maintenance of the Arctic Ocean sea ice cover have profound implications for the Late Tertiary climate of the Arctic and, more recently, this debate has acquired new significance in the context of possible climate effects of carbon dioxide-induced warming in high latitudes (*cf.* Barry 1989). Studies of the sediment record of the central Arctic Ocean have focused on the lithostratigraphy and micropaleontology (*cf.* Herman 1989). Two diverse interpretations of this record have been proposed. One theory proposes that the Arctic Ocean has had a permanent sea ice cover since initial cooling at 4 Ma (Clark 1982). Alternatively, Herman and Hopkins (1980) and Herman *et al.* (1989) distinguished three distinct, environmentally-controlled depositional sequences, using both lithological and faunal data. In their reconstruction, the Arctic Ocean, between 2.5 and 5 Ma, was cold but free from perennial sea ice. Small but significant amounts of ice-rafted sand grains indicate that glaciers existed in high latitudes at this time. A climatic threshold was crossed between 3 and 2.5 Ma with the development of a sharp, density-salinity stratified Arctic Ocean. The perennial Arctic sea ice cover originated *ca.* 0.7 Ma when water temperature dropped to present levels.

The views of Herman and Hopkins (1980) seem to be more in agreement with geologic evidence on land. Funder *et al.* (1985), Carter *et al.* (1986), Repenning *et al.* (1987) and Brigham-Grette and Carter (1992) all present data suggesting that the margins of the Arctic were at least seasonally ice-free in the late Pliocene. This evidence includes the fossil remains of sea otters in Alaska that were intolerant of perennial sea ice conditions, and high energy coastal deposits on northern Greenland, indicative of more enhanced wave action.

1.4.3 Early Pleistocene warm period

Raymo *et al.* (1990) suggested that the climatic deterioration of the late Pliocene was irregular rather than uni-directional. This point is suitably illustrated when one considers that the ecosystem of the Arctic Ocean borderlands experienced a complete climate cycle between 3 and 2 Ma (Repenning 1990). At 3 Ma, mixed coniferous forests grew along the margin of the Arctic Ocean whereas by 2.5 Ma, conditions had deteriorated to tundra and forest tundra with the first records of permafrost (Sher *et al.* 1979; Matthews and Ovenden 1990). Also, by *ca.* 2.5 Ma, the first regional Cenozoic Northern Hemisphere glaciation had begun (Shackleton and Opdyke 1977). Although the Arctic Ocean experienced a marked increase in ice rafting, it was at least seasonally ice-free during this period (Herman *et al.* 1989). By 2 Ma, sediments indicating ice-rafting along northern Greenland had been replaced by high energy coastal deposits containing a warmer marine fauna (Funder *et al.* 1985). Forest tundra, comparable to modern Labrador, grew in nearby sheltered sites at this time (Bennike and Böcher 1990). The Scandinavian ice sheet was greatly reduced by 2 Ma (Jansen *et al.* 1988; Jansen and Sjøholm 1991), and tundra was being replaced by forest vegetation in northeastern Siberia (Repenning 1984). Clark (1982) and Gilbert and Clark (1983) recognized somewhat more favourable conditions in the Arctic Ocean between 2 and 0.7 Ma.

One factor which may have caused this early Pleistocene warm period has been discussed by Clark *et al.* (1990). They report an abrupt and dramatic increase in calcareous species diversity and abundance in the central Arctic Ocean which may be explained by increased circulation with the North Greenland Sea. They postulate that accelerated spreading along the mid-ocean ridge through Fram Strait, facilitated improved circulation and the introduction of warmer, carbonate-rich water into the central Arctic Ocean within the past 2 Ma.

1.5 EARLIEST REGIONAL GLACIATION

The earliest **regional** glaciation of the Canadian Arctic likely occurred during the first Late Cenozoic glaciation of the Northern Hemisphere. Although Jansen and Sjøholm (1991) documented ice-rafted debris (IRD) in the Norwegian Sea over the last 5.5 Ma, they demonstrated that ice extent was much smaller before 2.5 Ma than after, as indicated by an increase in the IRD flux by 2 to 3 orders of magnitude. Similarly, Herman and Hopkins (1980, p. 209) reported small but significant amounts of IRD in central Arctic Ocean sediments between 4.5-5 Ma and 2.5 Ma, after which "the much greater abundance of coarse IRD...indicates a great increase in the frequency and intensity of glaciation on adjacent landmasses." These IRD records are consistent with other lines of evidence, such as the oxygen isotope record, which suggest that the first regional Late Cenozoic glaciation of the Northern Hemisphere occurred *ca.* 2.5 Ma (Shackleton *et al.* 1984; Carter *et al.* 1986; Repenning *et al.* 1987).

England (1987) suggested that if there was ever a regional ice sheet over Arctic Canada, that it occurred on a prefaulted, contiguous landmass during the late Tertiary. The absence of glacial geologic and geomorphic evidence beyond the last ice limit in central Ellesmere Island fiords prompted England (1987) to speculate that many fiords are not glacial, but tectonic in origin, and that they postdate pervasive glaciation of Arctic Canada. If faulting of a contiguous arctic landmass produced marine channels that subsequently prevented the buildup of a pervasive ice sheet, then perhaps the earliest regional glaciations were the only pervasive glaciation of the Canadian High Arctic. Because the Beaufort Formation was deposited on the margin of a contiguous arctic landmass and does not contain glacigenic sediments (Fyles 1990), then the earliest regional glaciation must be younger than *ca.* 3 Ma. This is consistent with age estimates from ocean sediment records (see above). Consequently, if England's (1987) hypothesis is correct, then the breakup of a contiguous arctic landmass must have occurred since the earliest regional glaciation, possibly 2.5 Ma.

1.6 PLIOCENE TRANSGRESSIONS

The earliest Late Tertiary Arctic marine transgression has been documented from the Pliocene Beaufort Formation on Meighen Island (Fyles *et al.* 1991). Here, fossiliferous mud is interbedded with fluvio-deltaic sand up to 100 m asl. The marine unit contains shells of both Pacific (*Mya truncata*, *Hiatella arctica*) and Arctic/Atlantic affinity (*Arctica islandica*), indicating that the deposit must closely postdate the submergence of Bering Strait, about 3.2 Ma (Fyles *et al.* 1991). This age estimate confirms earlier dates based on foraminifera (McNeil 1990), strontium ratios (Kaufman *et al.* 1990; Matthews and Ovenden 1990), and amino acid ratios (Brigham-Grette *et al.* 1987).

Brigham-Grette and Carter (1992) described three Pliocene marine transgressions from the Alaskan Coastal Plain (Fig. 1.1) and provided age estimates using a combination of amino acid ratios, paleomagnetism, and vertebrate and invertebrate paleontology. From oldest to youngest these are: Colvillian (40-60 m asl, 2.7-2.48 Ma); Bigbendian (35-60 m asl; *ca.* 2.48 Ma); and Fishcreekian (25-35 m asl, 2.48-2.14 Ma). Using amino acid geochemistry, these transgressions were roughly correlated to Beringian I, II, and III transgressions on the Seward Peninsula, Bering Strait (Kaufman *et al.* 1989; Brigham-Grette and Kaufman 1990; Kaufman 1992) but proved younger than the Beaufort marine event on Meighen Island (Table 1.1).

On northern Greenland, the Kap København Formation contains two distinct marine units (Fig. 1.1). The oldest unit, Member A, consists of laminated marine mud with scattered dropstones and lenses of diamict and is taken as evidence for an expanded Greenland Ice Sheet (Bennike and Böcher 1990). The presence of molluscan species with Pacific affinity requires that this member is younger than 3.2 Ma (opening of Bering Strait), but >2.4 Ma on the basis of foraminiferal data (McNeil 1990).

Table 1.1: Circumarctic correlations of Pliocene marine transgressions (modified from Brigham-Grette and Carter 1992)

Alaskan Arctic Coastal Plain	Seward Peninsula Bering Strait	Northern Greenland	Ellesmere Island	Meighen Island
EARLY PLEISTOCENE WARM PERIOD				
Fishcreekian 2.14 - 2.48 Ma	Beringian I	Kap København F^m Member B 2 Ma		
FIRST LATE CENOZOIC GLACIATION OF NORTHERN HEMISPHERE 2.4 - 2.5 Ma				
Bigbendian 2.48 Ma	Beringian II	?	?	
		Kap København F^m Member A 2.4 - 3.2 Ma	White Point marine unit 2.4 - 3.2 Ma	
Colvillian 2.48 - 2.7 Ma	Beringian III	?	?	
				Beaufort F^m marine unit *ca.* 3 Ma
OPENING OF THE BERING STRAIT 3.2 Ma				

Member **A** may correlate with either the Bigbendian or Colvillian transgressions (Table 1.1; Brigham-Grette and Carter 1992).

Member **B** is found up to 230 m asl and contains both nearshore and offshore sediments (Bennike and Böcher 1990). Paleontological and paleobotanical evidence (Repenning *et al.* 1987; Matthews and Ovenden 1990) suggests an age of about 2 Ma which is supported by paleomagnetic measurements (Funder *et al.* 1985). Brigham-Grette and Carter (1992) used amino acid ratios to suggest a correlation between Member **B** and the Fishcreekian transgression (Table 1.1).

Raised marine sediments at White Point, Ellesmere Island (Fig. 1.1) occur up to 122 m asl and contain *Hiatella arctica* shells (Pacific affinity) and *cibicides grossa* foraminifera which, according to McNeil (1990), suggest a late Pliocene age, between 3.2 and 2.4 Ma. Amino acid ratios on shells are significantly lower than those in Beaufort marine sediments and are likely correlative to either the Colvillian or Bigbendian transgressions (Table 1.1; Brigham-Grette and Carter 1992).

In summary, the oldest transgression is associated with the Beaufort Formation and occurred soon after 3.2 Ma. If this transgression predated the first regional Late Cenozoic glaciation (see previous section), then it may be related to tectonic and/or geoidal eustatic effects (*cf.* Mörner 1987). A eustatic high sea level has been recognized along the coastal regions of the North Atlantic between 3.5 and 3 Ma (Cronin and Dowsett 1990).

All three Pliocene marine units on the Alaskan Coastal Plain contain striated erratics, indicating that glaciers reached the Arctic Ocean during, or just prior to these transgressions. Both Bigbendian and Fishcreekian transgressions overlap with the earliest regional glaciation; however, only Fishcreekian sediments contain Canadian Shield erratics which were likely transported into this area by a continental ice mass. The presence of these erratics on the Arctic Coastal Plain signifies the breakup of the earliest regional ice sheet and the subsequent ice-rafting of debris in a pre-Pleistocene sea.

It is suggested here that Members **A** and **B** from the Kap København Formation may be tentatively correlated with the Bigbendian and Fishcreekian transgressions, respectively (similar to Brigham-Grette and Carter 1992). Member **A** was deposited during expansion of the Greenland Ice Sheet whereas Member **B** is associated with a forest tundra flora and a warmer Arctic Ocean (Funder *et al.* 1985; Bennike and Böcher 1990). The magnitude of the relative sea level change (230 m asl) represented by Member **B** is similar to the maximum elevation (285 m asl) of raised marine deposits on northeastern Ellesmere Island. The latter transgression is associated with the retreat of the maximum Greenland ice advance onto Ellesmere Island (England *et al.* 1981).

If the marine transgressions on northeastern Ellesmere Island and northern Greenland are correlative, then: (1) Member **B** of the Kap København Formation represents a relative sea level adjustment associated with retreat of the Greenland Ice Sheet; (2) the age of the 285 m shoreline on northeastern Ellesmere Island has been underestimated (0.5-1 Ma based on amino acid analysis, England 1987) and 2 Ma would be a closer approximation (proposed age of Member **B**); and (3) the maximum glaciation documented on northeastern Ellesmere Island occurred during the first Late Cenozoic glaciation of the Northern Hemisphere possibly at 2.5 Ma.

1.7 PRE-WISCONSINAN GLACIATIONS

Evidence for pre-Wisconsinan glaciations of the Canadian Arctic islands consists predominantly of scattered observations on meltwater channels, erratics, and isolated raised marine sediments beyond the last ice limit. Although sparse, this evidence demonstrates that glaciers were most extensive prior to the last glaciation.

Evidence for multiple glaciations and transgressions have been observed along the eastern side of Nares Strait, a 20-60 km wide channel separating Ellesmere Island from Greenland (Christie 1967; England 1978; England and Bradley 1978; England *et al.* 1978, 1981; Retelle 1986; Lemmen and England 1992). The oldest glaciation involved the advance of the northwest Greenland Ice Sheet at least 20 km onto northeast Ellesmere Island where distinctive crystalline erratics outline its maximum limit. Local summits lying above these erratics were likely never glaciated (England 1978). Outlet glaciers from local ice caps on Ellesmere Island abutted Greenland ice in tributary valleys (Retelle 1986; Lemmen and England 1992). Initial Greenland ice retreat was associated with ice-dammed lakes and a relative sea level at 285 m asl. Later, in response to further Greenland ice retreat, Ellesmere ice advanced through tributary valleys to Nares Strait where it was forced to float as a series of small ice shelves in glacioisostatically depressed embayments (England *et al.* 1981; Lemmen and England 1992). The relative sea level at this time was 175 m asl.

Shells transported by Greenland ice onto Ellesmere Island may date ≥1 Ma and provide a maximum age for this glaciation (England 1987). Retelle (1986) proposed a minimum age of >70 000 BP for the Robeson Aminozone which included all samples related to the retreat of Greenland ice from northeast Ellesmere Island. Lemmen and England (1992) concluded that climatic factors may have been most favourable for extensive glaciation on northern Ellesmere Island when Arctic Ocean sea ice was reduced, possibly between 0.4 and 0.6 Ma.

1.8 WISCONSINAN GLACIATION

Research on the Late Pleistocene glacial history of the Arctic Archipelago has been dominated by the debate of maximum versus minimum ice sheet coverage (*cf.* England 1974, 1976*a*; Paterson 1977; Hodgson 1989; Prest 1990). Craig and Fyles (1960) portrayed an **Ellesmere-Baffin Glacier Complex** contiguous with the northeastern part of the Laurentide Ice Sheet during the Wisconsin Glaciation (Fig. 1.5). Islands farther west supported only local ice caps or remained ice-free. It has since been shown that the Ellesmere-Devon Island ice cover was quite separate from Baffin Island ice which was part of the Laurentide Ice Sheet (Prest 1970). However, the concept of an Ellesmere-Baffin glacier complex resurfaced with the publication of Flint's (1971) sketch map showing "all-time glacier cover" on North America and Greenland.

REMOVED DUE TO COPYRIGHT RESTRICTIONS

Figure 1.5: The Ellesmere-Baffin Glacier Complex (from Craig and Fyles 1960).

Blake (1970) proposed a Late Wisconsinan **Innuitian Ice Sheet** situated over the entire Queen Elizabeth Islands to account for an apparent ridge of greatest postglacial emergence through the central Queen Elizabeth Islands (Fig. 1.6). This ridge

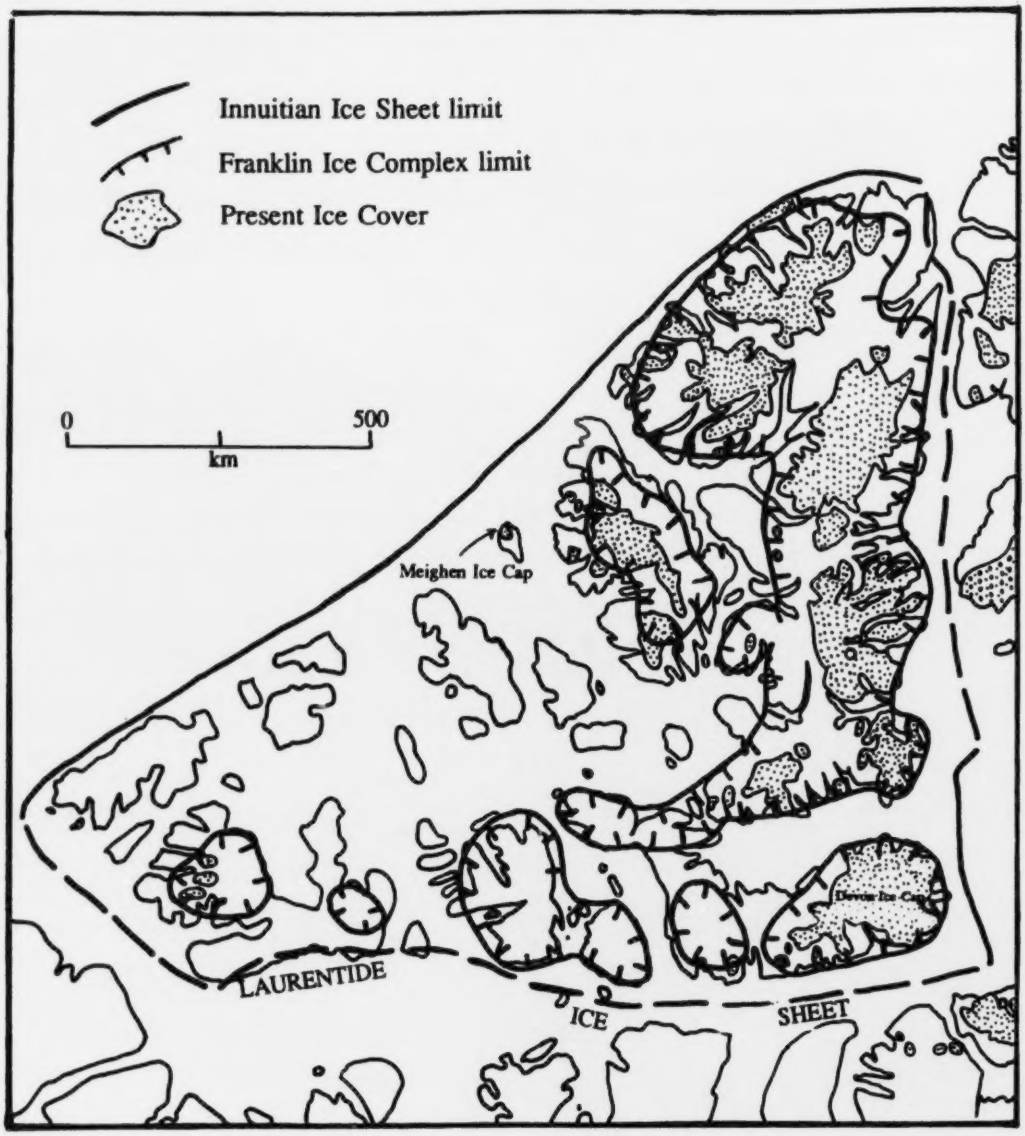

Innuitian Ice Sheet limit
Franklin Ice Complex limit
Present Ice Cover
0
500
km
Meighen Ice Cap
Devon Ice Cap
LAURENTIDE
ICE
SHEET

Figure 1.6: Maximum (Innuitian Ice Sheet?) and minimum (Franklin Ice Complex?) extent of the last glaciation on the Queen Elizabeth Islands (modified from Hodgson 1989).

coincided with the main divide of Fortier and Morley's (1956) submerged and glacially modified drainage system. Furthermore, a pervasive ice sheet was consistent with the interpretation that glacially overdeepened channels existed between the western and central Queen Elizabeth Islands (Pelletier 1966; Horn 1963; Marlowe 1968; Hattersley-Smith 1969). The demise of the Innuitian Ice Sheet was apparently recorded by the sequential penetration of driftwood and shells into fiords and marine channels.

The Innuitian Ice Sheet has remained in the literature for two decades despite numerous field observations that obviate its physical basis. As early as 1965, Fyles remarked that erratics associated with the major ice cover of the western islands were of Laurentide origin, a conclusion supported by Hodgson (1989, 1990). England (1976) reinterpreted the two main criteria upon which the Innuitian Ice Sheet was based (the pattern of emergence and the entry of shells and driftwood). He also incorporated newer fieldwork and paleoclimatic theory to reject the ice sheet during the last glaciation, replacing it with the non-contiguous Franklin Ice Complex (Fig. 1.6). England (1983, 1990) provided evidence that, distal to the Franklin Ice Complex, the fiords and channels were occupied by a full glacial Innuitian Sea. Evans (1990) provided evidence of such full glacial marine fauna between 10 000 and 20 000 BP in fiords of northwestern Ellesmere Island, demonstrating that a submerged peripheral depression existed around the last ice margin. Lemmen (1989) also showed that terrestrial bryophytes occupied ice-free areas at 14 000-16 000 BP.

Some authors continue to propose models of all-pervasive ice sheets in the high Arctic for the last glaciation (CLIMAP 1976; Hughes *et al.* 1977, 1985; Denton and Hughes 1981; Mayewski *et al.* 1981; Reeh 1984; Tushingham and Peltier 1991). Most recently, Tushingham (1991) revived the concept of the Innuitian Ice Sheet on the basis of a geophysical model which utilizes the global relative sea level record to reconstruct former ice sheets. He argued that the magnitude of postglacial emergence observed in the high arctic, which to date has proved enigmatic with regard to a limited last ice load, required the former presence of a pervasive ice sheet, retreating at 18 000 BP. The absence of glacial geologic and geomorphic evidence supporting such a deglacial model is dismissed on the grounds that the ice sheet was cold-based and consequently, left no evidence. This argument has been repeatedly rejected (*cf.* England 1987; England *et al.* 1991) and it has been emphasized that Tushingham (1991) ignored the wealth of geological data which demonstrate that deglaciation from the last ice limit began 8000-10 000 BP. Moreover, the geophysical model provides a better fit to the observed emergence data if a limited ice load, retreating at 8000-10 000 BP, is used as input rather than a pervasive Innuitian Ice Sheet retreating at 18 000 BP (England *et al.* 1991).

During the last decade, many local studies on northern and central Ellesmere Island and Greenland have provided the necessary physical evidence to firmly establish the limits of the Franklin Ice Complex, and to describe the style and timing of deglaciation (England 1978, 1983, 1985, 1990, 1992; England and Bednarski 1986; Bednarski 1986; Retelle 1986; Lemmen 1989; Evans 1990; Sloan 1990). This evidence can be summarized as follows.

1.8.1 Ice buildup

Details on ice buildup are uncertain. England (1985, 1990) previously suggested that pre-Holocene shells from the submerged peripheral depression indicate that ice buildup and associated isostatic loading had commenced prior to 30 000 BP. Lemmen (1989) argued that the last ice maximum on Marvin Peninsula, northernmost Ellesmere Island was likely achieved as late as 11 000 BP, based on the youngest date on glacially transported organic remains. Furthermore, Lemmen and England (1992) suggested that the last ice limit was more extensive than any earlier Wisconsinan advance. This is supported by Evans' (1990) description of a possible Sangamonian peat below a till from the last ice advance on northern Ellesmere Island. Most recently, England (1992) has favoured a relative sea level history which combines a rapid buildup of ice after 14 000 BP with global eustatic sea level rise, causing a rapid transgression to marine limit by *ca.* 9000-10 000 BP.

1.8.2 "Full glacial sea"

The concept of the "full glacial sea" was introduced by England (1983) to describe the sea rather than glacier ice occupying the fiords and marine channels of northeastern Ellesmere Island during the last glaciation. In this model, sea level was originally considered to be stable from at least 11 000 until 8000 BP, recording full glacioisostatic loading that was apparently in isostatic equilibrium (England 1983, p.908).

> "Because no glacial unloading, and consequently no rebound, occurred between 11-8 ka BP, the stability of sea level during that interval cannot be explained as a balancing of rebound and sea level rise. Assuming that local eustatic changes were zero[1] between 11 and 8 ka BP then sea level stability must indicate that the ice load was in isostatic equilibrium."

England (1990) extended the limits of the "full glacial sea" to west-central Ellesmere Island and named it the **Innuitian Sea**. With the realization that the relative sea level history of the high arctic must contain a glacioeustatic component, England (1992) revised the concept of the Innuitian Sea. In many ways these revisions follow an

---

[1] The assumption that postglacial eustatic sea level change in high latitudes was zero originated from a diagram in Clark *et al.* (1978) where global eustatic sea level change was predicted using a viscoelastic sea level model. The possibility that the eustatic change could even be negative since 16 000 BP in the High Arctic was suggested informally by one of the authors (J. England, pers. comm. 1992).

REMOVED DUE TO COPYRIGHT RESTRICTIONS

Figure 1.7: Generalized limit of last glaciation on northern and western Ellesmere Island. See text for discussion.

earlier proposal by Hodgson (1985, p. 366) in which he suggested that "a shallower glacial sea deepened during a worldwide postglacial sea-level rise, perhaps aided by a local early Holocene ice-cap expansion." The Innuitian Sea does not refer to a particular sea level during the last glaciation for it would have existed at any range of elevations from below present sea level up to the marine limit. Because the Innuitian Sea occupied the peripheral depression beyond the last ice limit, it must record the complete loading and unloading history associated with the last glaciation; hence, it is of plaeoclimatic and geophysical interest (England 1983, 1992). The Innuitian Sea must have been transgressive between 18 000 and 8800 BP because eustatic sea level rise began at 18 000 BP (Fairbanks 1989) and exceeded uplift until 8800 BP (England 1992). A period of apparent sea level stability between 8800 and 8000 BP requires a balance between the rate of uplift and the rate of eustatic sea level rise in order to sustain it (England 1992).

Isobases drawn on the limit of the Innuitian Sea reflect the pattern of glacial loading across northern and central Ellesmere Island and Greenland. In their reconstructions, England and Bednarski (1986), Evans (1990) and Lemmen (1988) demonstrated that isobases on the 10 000 BP shoreline mimic the loading centres over the Grant Land Mountains and the Greenland Ice Sheet, with an intervening low cell over the Hazen Plateau (Fig. 1.8). From these isobase maps it was also apparent that the Greenland loading centre glacioisostatically influenced the northeastern coast of Ellesmere Island (England 1976*b*, 1983; Retelle *et al.* 1989). England (1992) presented the isobase pattern for the 8000 BP shoreline in Greely and Cañon fiords which he interpreted to be the upper limit of the Innuitian Sea (Fig. 1.9). In this area, the isobases apparently portray a narrow plunging ridge which, according to England (1992), is unlike isobases drawn for other glaciated areas of arctic Canada where broad cells of emergence typify former ice dispersal centres. England (1992) offered the possibility that the 8000 BP isobase pattern may contain a neotectonic component or reflect structurally-controlled glacioisostatic emergence. The ridge parallels the structural trends.

REMOVED DUE TO COPYRIGHT REASONS

Figure 1.8: Isobases drawn on the 10 000 BP shoreline on northeastern Ellesmere Island and northwest Greenland. The isobase pattern shows the relative magnitude of the last ice load and the limit of the Innuitian Sea (from England and Bednarski 1986).

1.8.3 Deglaciation

The last glacial maximum attained by the Franklin Ice Complex represents a 5-60 km advance of existing upland icefields and, in the case of the Agassiz Ice Cap, a 200 m

REMOVED DUE TO COPYRIGHT REASONS

Figure 1.9: Limit of the last glacial maximum in Greely and Cañon fiords. Isobases are drawn on the limit of the Innuitian Sea (from England 1992).

increase in thickness under the ice divide (Koerner *et al.* 1987). Plateau ice caps also formed on surfaces now ice-free (≥400 m asl). Glacier retreat on northwest Ellesmere Island was underway by 10 000 BP and, following a pronounced calving phase, some ice limits lay within 500 m of present margins by 8300 BP (Evans 1990). Bednarski (1986) and Lemmen (1989) reported slow ice retreat along the north coast between 10 000 and 8000 BP but rapid deglaciation thereafter. In contrast, the area south of the Grant Land Mountains on northeast Ellesmere Island experienced slow initial deglaciation between 8000 and 6000 BP, followed by steady, rapid retreat during the mid Holocene (England 1983). England (1985) also reported initial ice retreat from the last glacial maximum on Hall Land, Northwest Greenland, at 8000-8200 BP which agrees with the onset of initial emergence on northeast Ellesmere Island. Retelle (1986) recorded the retreat of a local ice cap farther north which contacted the sea along Nares Strait. The oldest deglacial date of 7500 BP is a minimum estimate on initial ice retreat and compares favourably with England's observations to the southwest.

In Greely and Cañon fiords, west-central Ellesmere Island, England (1990) and Sloan (1990) reported initial deglaciation between 8000 and 7800 BP with some glaciers back at their present margins before 7500 BP. England (1992) attributed this rapid deglaciation to the effect of eustatic sea level rise on the calving rates of tidewater glaciers. Likewise, he considered that the pronounced calving phase that led to rapid ice retreat between 10 000 and 8000 BP on northwestern Ellesmere Island (Evans 1990) was also triggered by a eustatic sea level rise. Having established that this deglacial mechanism was common during initial retreat from the last ice limit, England (1992) reasoned that it may also explain why glaciers on opposite sides of the Grant Land Mountains retreated 2000 years out of phase. England (1992) concluded that eustatic sea level rise would have caused increased calving of tidewater glaciers in the northern fiords, whereas those glaciers that terminated on the Hazen Plateau were unaffected by sea level change. This apparent regional disparity had been previously attributed to variations in topoclimate and ice dynamics, in addition to glacioclimatic controls (England and Bednarski 1986; Lemmen 1989).

Hodgson (1985) proposed three possible models for deglaciation of fiords opening into Eureka Sound. Model A describes the breakup of a pan-archipelago Innuitian Ice Sheet in order to explain the proposed pattern of glacioisostatic unloading. Model **B** places the limit of the last ice at the fiord heads with the Innuitian Sea occupying the fiords (England 1983, 1990). Retreat from the fiord heads began 8000-7500 BP with ice fronts well upvalley by 7000 BP. Model **C** proposes that initial deglaciation was under way by 9000 BP followed by a subsequent rapid readvance of partially floating ice to at least the middle fiords, and final retreat at 7500 BP.

1.8.4 Postglacial emergence

Isobases drawn on the 8000 and 6000 BP shorelines along northern Ellesmere Island outline the pattern of postglacial unloading as the ice cover on the Grant Land Mountains slowly retreated at 10 000 BP, and rapidly diminished after 8000 BP (Bednarski 1986; England and Bednarski 1986; Evans 1990; Lemmen 1988). Meanwhile, emergence on northeastern Ellesmere Island was primarily governed by the northwest Greenland ice load which remained relatively stable between 10 000 and 8000 BP (England 1985). As a result, no emergence is observed along northern Nares Strait until after 8000 BP. Subsequently, the northwest margin of the Greenland Ice Sheet retreated, causing moderate emergence along the adjacent Ellesmere Island coast (Retelle *et al.* 1989), despite the fact that most glaciers on the Hazen Plateau remained within a few kilometres of their limit for a further 2000 years (England 1983).

In Greely and Cañon fiords, west-central Ellesmere Island, England (1992) reported a stable sea level between 8800 and 8000 BP which he attributed to a balance between uplift and eustatic sea level rise (B-C, Fig. 1.10). This balance continued between 8000 and 7800 BP when ice margins retreated without an observable decrease in the elevation of marine limit as would be expected (restrained rebound; *cf.* Andrews 1970). Between 7800 and 7200 BP, emergence was observable but slow, even though many glaciers had retreated to their present margins (C-D, Fig. 1.10). This is attributed to rapid calving which would have had a negligible effect on the overall glacioisostatic load. After 7200 BP, rapid emergence occurred in response to rapid glacial unloading (D-E, Fig. 1.10).

REMOVED DUE TO COPYRIGHT RESTRICTIONS

Figure 1.10: Provisional emergence curve for inner Greely Fiord (from England 1992).

1.9 HOLOCENE ENVIRONMENTS

Bradley (1990) provided an extensive review of the Holocene paleoenvironmental record of the Canadian High Arctic using proxy data primarily from ice cores, lake sediment cores, surface peat accumulations, dated driftwood occurrence, and the marine faunal record. Only the main departures of the record are outlined here. High arctic ice cores show a continuously warm period from 10 000 to about 4000 BP with the warmest period on the Agassiz Ice Cap between 9000 and 7000 BP (Fisher *et al.* 1983; Koerner and Fisher 1990) and on the Devon Ice Cap about 5000 BP (Koerner 1977; Paterson *et al.* 1977; Fisher and Koerner 1981). Blake (1972) reported maximum driftwood penetration of interisland channels between 6500 and 4500 BP, and Stewart and England (1983) corroborated this on the north coast of Ellesmere Island with the earliest driftwood entry extending back to 8900 BP. Evans (1989) documented the former occurrence of narwhal bordering the Arctic Ocean during the early Holocene (6800 BP), and Bednarski (1990) dated the remains of a bowhead whale at 7500 BP in Nansen Sound. Both specimens represent an early Holocene range extension of 400-700 km and suggest that sea ice and ice shelf conditions were less severe at that time.

Ice cap records indicate a cooling trend over the last 4000-5000 years although there have been intermittent periods of climatic amelioration. The inception of Meighen Ice Cap (Koerner 1968) and the overriding of mid Holocene organics by advancing glaciers were a direct response to this cooling trend (Dyck and Fyles 1963; Long 1967; Lowdon and Blake 1979; Blake 1981). Ice shelves regrew along the northern coast of Ellesmere Island and heavier sea ice conditions excluded driftwood from the fiords and marine channels (Lyons and Mielke 1973; Lemmen *et al.* 1988). Temperatures reached a minimum between 400 and 100 BP, perhaps the coldest period of the entire Holocene. Many glaciers advanced to their postglacial maximum positions at this time (Müller 1964; Hattersley-Smith 1972; Blake 1981; King 1983; Bergsma *et al.* 1984; Evans 1988).

A marked temperature increase has been observed since 1925 and, consequently, glaciers and ice caps throughout the archipelago have experienced negative mass balances and ice recession. Bradley (1990) compared this exceptionally warm period to the climate conditions of the early to mid Holocene.

CHAPTER TWO

Study area, research objectives and methods

This chapter introduces the physical geography of western Fosheim Peninsula and summarizes previous Quaternary research in the region. The objectives and rationale for this study are outlined within the context of prevailing models of glaciation and relative sea level change and our current knowledge on landscape evolution in the Canadian High Arctic.

2.1 ORIENTATION

Fosheim Peninsula constitutes part of the Eureka Sound lowlands which are bordered by mountain ranges and associated ice caps (Fig. 2.1). These include; the Grant Land Mountains and ice cap (2500 m) to the north, the Victoria and Albert Mountains and Agassiz Ice Cap (2000 m) to the east, the Prince of Wales Mountains and Central Ellesmere ice cap (2400 m) to the southeast, and the Princess Margaret Range (Axel Heiberg Island) and Müller Ice Cap (2200 m) to the west (Fig. 2.1).

The Eureka Sound lowlands are dissected by three main fiords and interisland channels (Fig. 2.1). Greely Fiord - Nansen Sound is the largest inlet in the Queen Elizabeth Islands, up to 30 km wide, 400 km long, and 920 m deep (Hattersley-Smith 1969; England 1987). Eureka Sound is a narrow, winding channel, between 10 and 20 km wide and 300 km long, separating Ellesmere Island from Axel Heiberg Island. The channel diverges around Stor Island at the mouth of Bay Fiord (Fig. 2.1). Water depths increase northwards from 200-300 m in central Eureka Sound to 400-550 m where it enters into the deepest basin of Greely Fiord (>800 m deep). Cañon fiord measures 110 km long and up to 13 km wide; however, no bathymetric data are available. Both Cañon and Bay fiords penetrate the central region of Ellesmere Island and act as major drainage outlets for the Agassiz and Central Ellesmere ice caps (Fig. 2.1).

The study area in western Fosheim Peninsula is *ca.* 5500 km^2. It is bordered on three sides by Eureka Sound, Greely Fiord and Cañon Fiord and is isolated from the

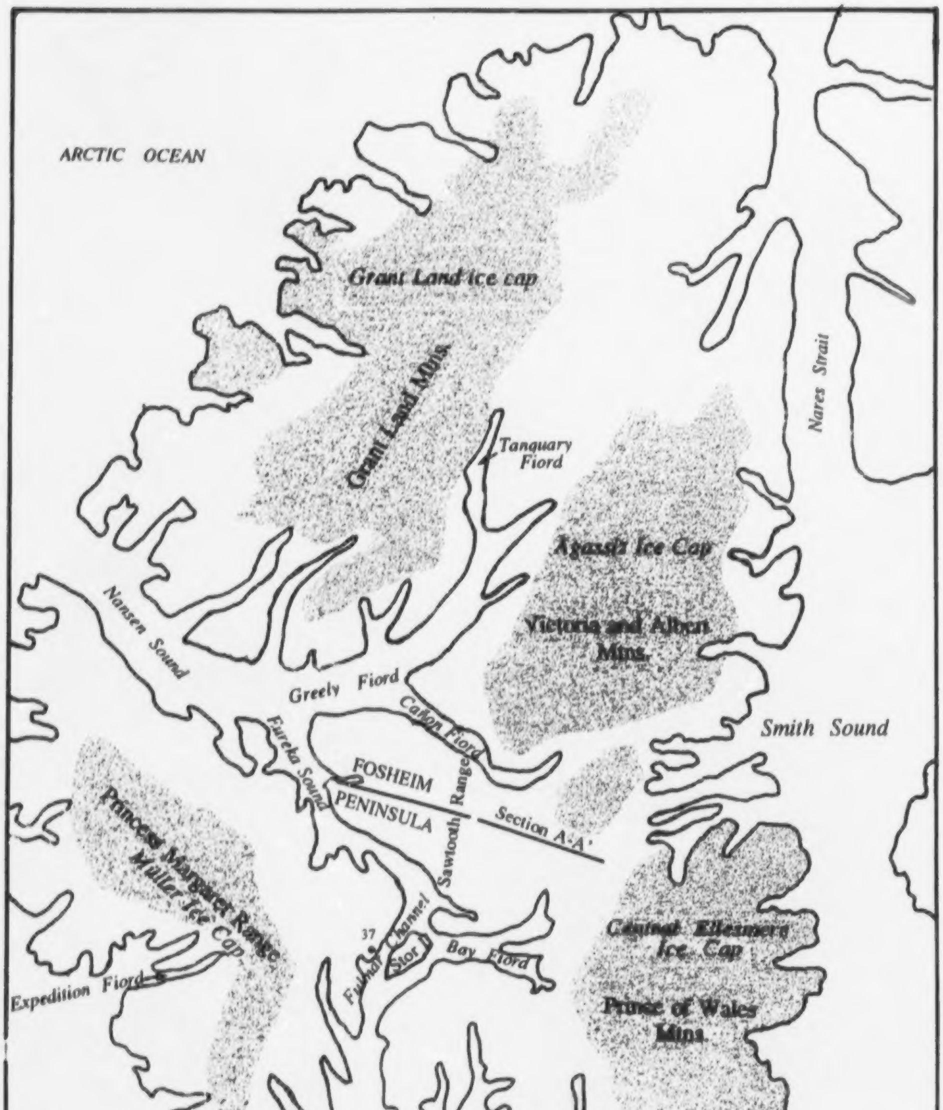

ARCTIC OCEAN
Grant Land ice cap
Grant Land Mtns.
Tanquary Fiord
Nares Strait
Agassiz Ice Cap
Victoria and Albert Mtns.
Nansen Sound
Greely Fiord
Cañon Fiord
Smith Sound
Eureka Sound
FOSHEIM PENINSULA
Sawtooth Range
Section A-A'
Princess Margaret Range
Müller Ice Cap
Expedition Fiord
37
Fulmar Channel
Stor Ø
Bay Fiord
Central Ellesmere Ice Cap
Prince of Wales Mtns.
Norwegian Bay
Graham I.
Sydkap Ice Cap
0
200
km

Figure 2.1: Eastern Queen Elizabeth Islands showing location of western Fosheim Peninsula. Note location of geologic cross-section in Figure 2.5.

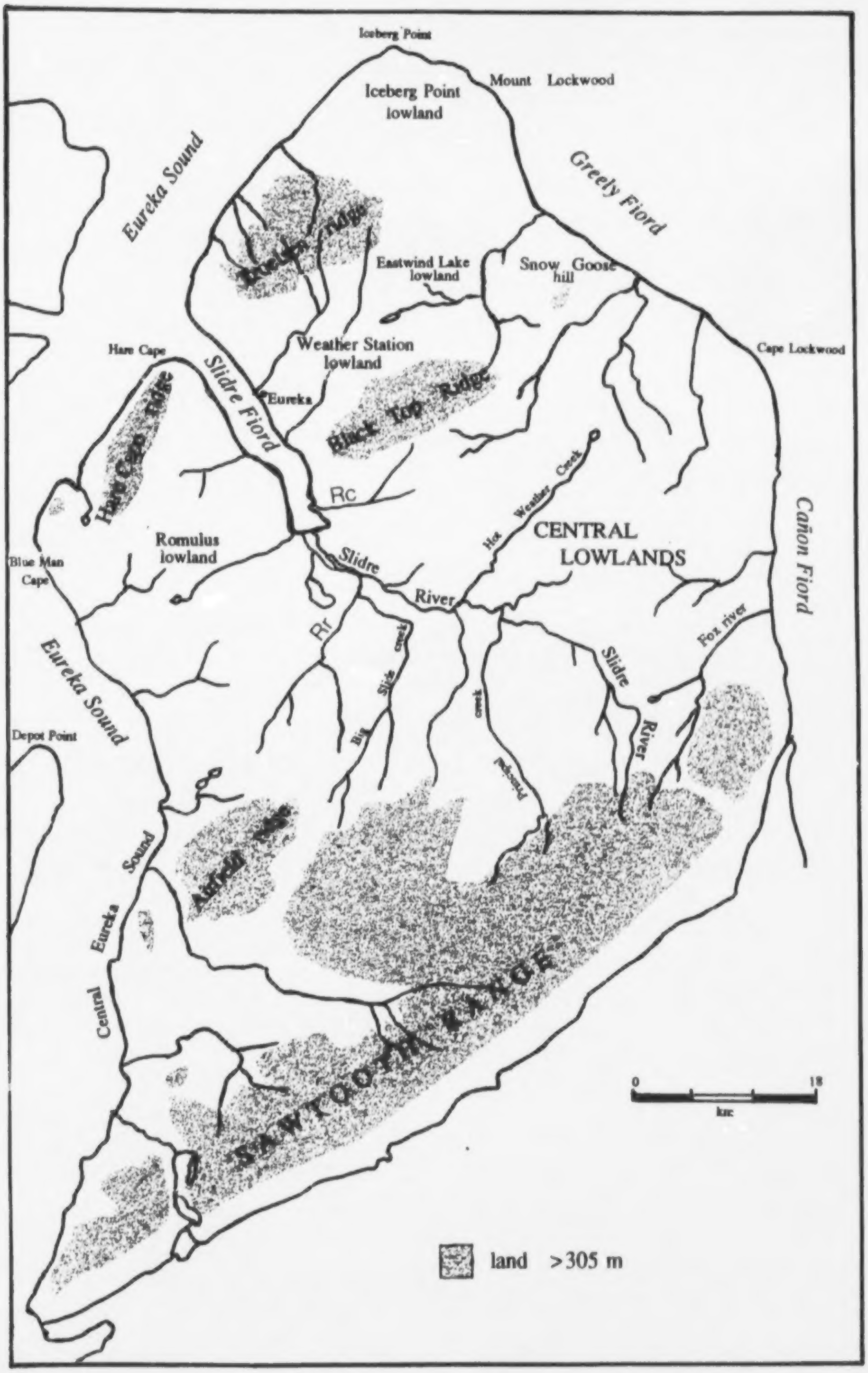

Iceberg Point
Mount Lockwood
Iceberg Point lowland
Eureka Sound
Greely Fiord
ridge
Eastwind Lake lowland
Snow Goose hill
Hare Cape
Cape Lockwood
Weather Station lowland
Slidre Fiord
Eureka
Black Top Ridge
Hare Cape ridge
Rc
Romulus lowland
Hot Weather Creek
CENTRAL LOWLANDS
Blue Man Cape
Slidre
River
Cañon Fiord
Rr
Fox river
Eureka Sound
Big Slide creek
creek
Slidre
Principal
River
Depot Point
ridge
Central Eureka Sound
SAWTOOTH RANGE
0
18
km
▨ land >305 m

Figure 2.2: Official and unofficial placenames of western Fosheim Peninsula. Rc = Remus Creek, Rr = Romulus river. See text for details.

remainder of the peninsula by the Sawtooth Range (Fig. 2.2). This range contains the highest peak (1026 m) and the only glaciers in the study area. Other upland areas include Black Top Ridge (838 m), Hare Cape ridge* (627 m), Troelsen ridge* (661 m), Airfield ridge* (692 m), and Snow Goose hill* (288 m). Slidre River and its main tributaries (Remus Creek, Hot Weather Creek, Principal creek*, and Big Slide creek*) drain the interior of the peninsula and flow into the head of Slidre Fiord. Numerous other small streams exist on the peninsula but are unnamed.

Inuit have occasionally occupied Fosheim Peninsula as evidenced by stone tent rings and caches located on the shores of Slidre Fiord and Eureka Sound (Fig. 2.3). These were likely seasonal hunting camps used by the Thule culture (P. Schledermann, pers. comm. 1989). The first European explorers to visit the region were Sverdrup and Fosheim during the *Second Norwegian Polar Expedition* of 1898-1902. Together they travelled by dogsled along Eureka Sound to Slidre Fiord where Fosheim headed east to investigate the peninsula and Sverdrup continued north into Greely Fiord and Nansen Sound (Sverdrup 1904).

Eureka weather station, operated by the Atmospheric Environment Service, is situated at the mouth of Station Creek on the northern shore of Slidre Fiord (Figs. 2.2 and 2.4). It was established in April 1947 as part of the *Joint Arctic Weather Stations* that involved both Canada and the United States. Since then, the station has played a major role in the promotion of scientific research in the region. In 1952 Troelsen used the weather station as a base to make some of the first observations on the glacial history and geological evolution of the region (Troelsen 1952). During *Operation Franklin* (1955), McMillan, Sim and Marsden travelled from Eureka to Slidre River valley, Hare Cape ridge and the Sawtooth Range on scientific investigations (Sim 1961; McMillan 1963; M. Marsden, pers. comm. 1990). The first oil discovery in the Arctic Islands was made near Eureka in 1972 and although the reserves remain untapped, planning for the export of oil and gas by pipeline prompted extensive mapping of surface materials and ground ice in west-central Ellesmere Island (Hodgson 1974; Hodgson and Edlund 1977). More recently in 1988, the Geological Survey of Canada established a Global Change observatory at Hot Weather Creek to monitor the effects of climate change on high arctic environments. This has resulted in a marked increase in the amount and diversity of research activity on the peninsula.

2.2 GEOLOGY

A structural cross-section from northern Eureka Sound to Smith Sound provides a regional framework for the bedrock geology of the study area (Fig. 2.5). The

---

* denotes unofficial name for use in this thesis. It appears at the first reference only.

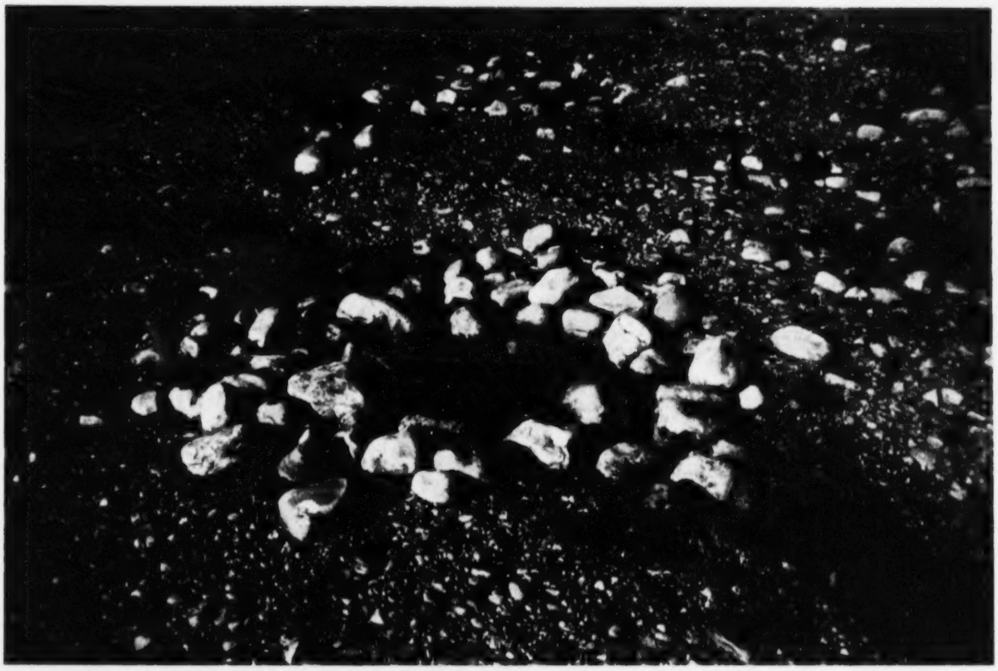


Figure 2.3: Stone cache and nearby tent rings from a seasonal hunting camp built by Thule Inuit. The camp is located on Romulus moraine at 129 m asl and *ca.*10 km from Eureka Sound.

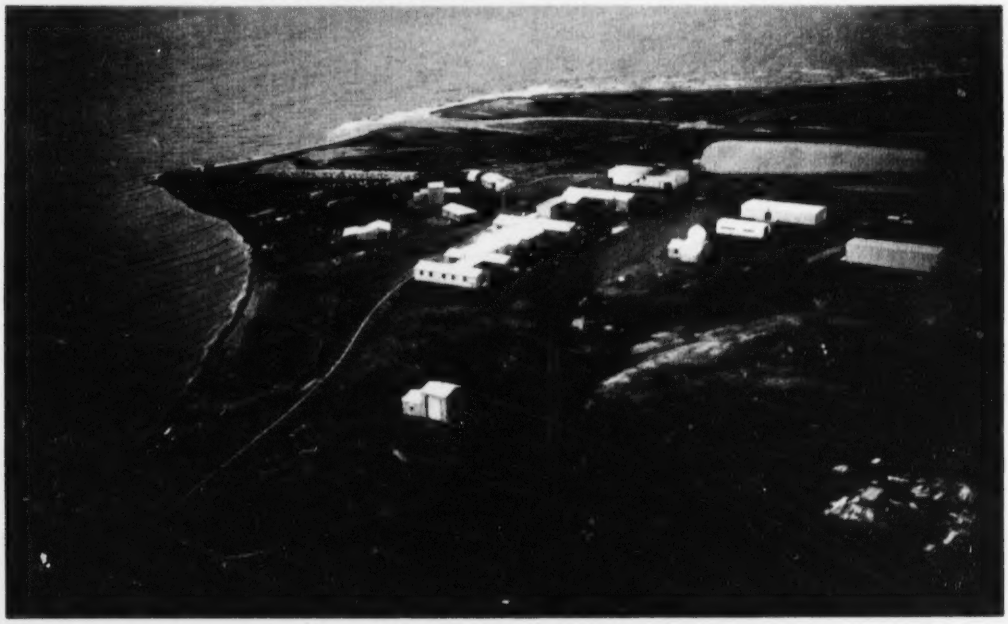


Figure 2.4: Eureka weather station on the northern shore of Slidre Fiord. Since its establishment in 1947, researchers have enjoyed the hospitality and facilities (*e.g.* the bar!) at the station.

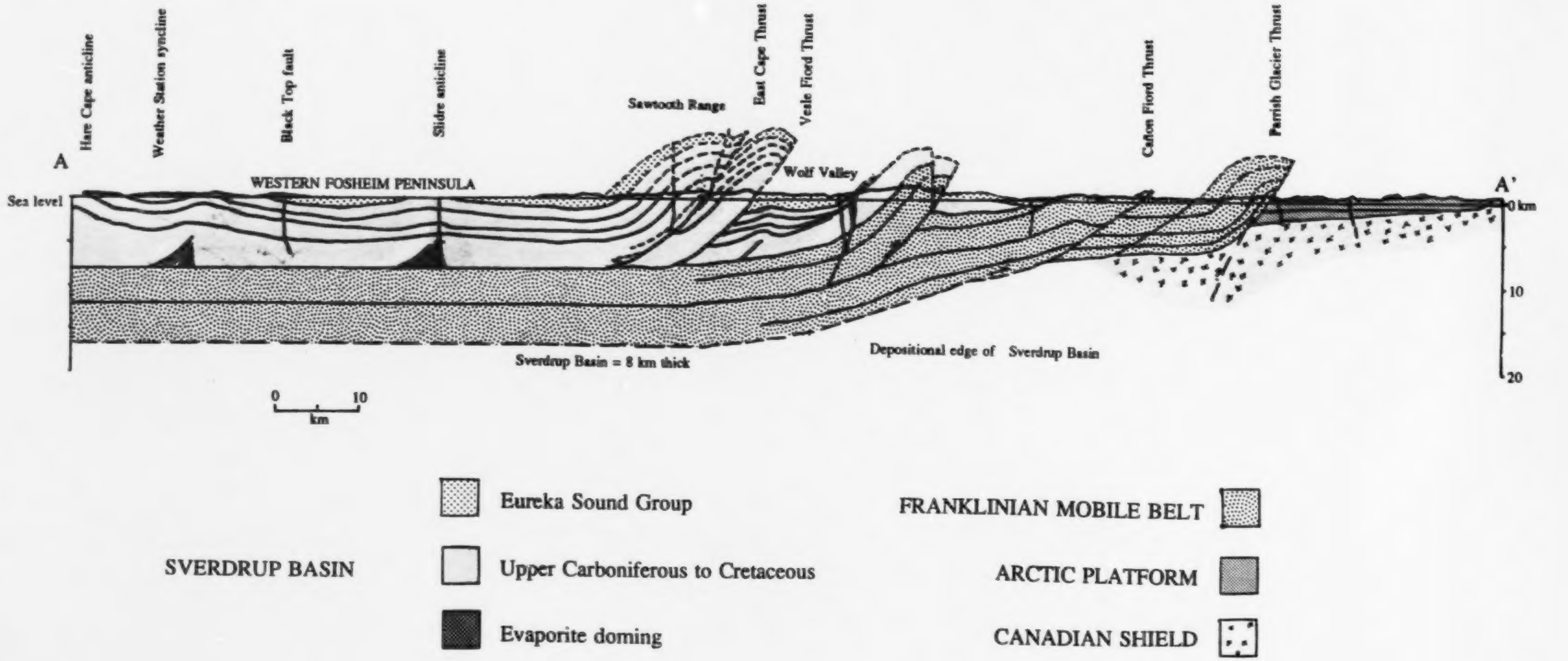

Hare Cape anticline
Weather Station syncline
Black Top fault
Slidre anticline
Sawtooth Range
East Cape Thrust
Veale Fiord Thrust
Wolf Valley
Cañon Fiord Thrust
Parrish Glacier Thrust
A
A'
Sea level
WESTERN FOSHEIM PENINSULA
0 km
10
20
Sverdrup Basin = 8 km thick
Depositional edge of Sverdrup Basin
0
10
km
SVERDRUP BASIN
Eureka Sound Group
Upper Carboniferous to Cretaceous
Evaporite doming
FRANKLINIAN MOBILE BELT
ARCTIC PLATFORM
CANADIAN SHIELD

Figure 2.5: Structural cross-section AA' through Fosheim Peninsula. No vertical exaggeration. See Figures 2.1 and 2.6 for location. Modified from Ricketts (1987).

Fosheim Peninsula is underlain by folded and faulted sedimentary rocks of the **Sverdrup Basin** and **Franklinian Mobile Belt** (Thorsteinsson and Tozer 1970; Trettin 1989). The latter constitutes a narrow band along the southwestern margin of the peninsula which comprises part of the more regionally extensive Central Ellesmere Fold Belt (Fig. 1.3). Farther southeast, undeformed lower Paleozoic strata of the **Arctic Platform** pinch out against the Precambrian crystalline basement of the **Canadian Shield.**

Surface exposures in the study area consist almost entirely of Mesozoic and Cenozoic sedimentary rocks deposited in the Sverdrup Basin (Figs. 1.3 and 2.6). Episodic uplift of the adjacent craton and changes in relative sea level during the Mesozoic led to fluctuations in sediment supply and depositional environments such that either sandstones or shale-siltstones dominate. Mesozoic bedrock outcrops on the western margin of the peninsula and in the Sawtooth Range where, for the most part, it is represented by the Triassic Heiberg Formation and undivided Jurassic formations. These are dominantly fine to medium-grained, quartzose sandstone with interbedded siltstone and shale. Where beds are steeply dipping, the more resistant sandstone weathers to form strike-aligned ridges (Fig. 2.7). Mesozoic outcrops also include the shale and silty shale of the Jurassic Deer Bay and Triassic Blaa Mountain formations. Gabbro and diabase sills and dykes are particularly common in the Heiberg and Blaa Mountain formations.

The earliest phase of Eurekan tectonism caused the breakup of Sverdrup Basin into a series of smaller sub-basins. The **Fosheim Basin**, occupied most of Axel Heiberg and west-central Ellesmere islands (Ricketts 1987*c*, 1991). Sediments of the Eureka Sound Group within Fosheim Basin are more than 3 km thick and consist of three lithostratigraphic units (Ricketts 1988). The oldest is the *Expedition Formation* (middle/late Campanian to early Paleocene) which is composed predominantly of sandstone and minor shale. The *Strand Bay Formation* (early Paleocene) is mostly shale representing a basin-wide marine transgression; however, the sandstone component increases towards Fosheim Peninsula where coastal sedimentary environments dominated. Continued progradation gave rise to the overlying sandstone of the *Iceberg Bay Formation* (middle/late Paleocene to middle Eocene) which is the most commonly exposed geological unit on Fosheim Peninsula. It consists predominantly of fine- to coarse-grained, poorly cemented, quartzose sandstone with interbeds of shale, siltstone and coal (Fig. 2.8). Coal-bearing strata may include petrified logs and stumps indicating a broad-leaved deciduous forest (Basinger 1991). Farther south, vertebrate fossils such as tortoises, lizards, crocodiles and lemurs have been recovered from carbonaceous beds near the top of the formation (West *et al.* 1977).

The Iceberg Bay Formation represents the final depositional event of the Fosheim Basin prior to its uplift and thrusting during the Eurekan Orogeny. Fosheim Basin was partially rejuvenated during the Miocene and/or Pliocene when fluvial sediments

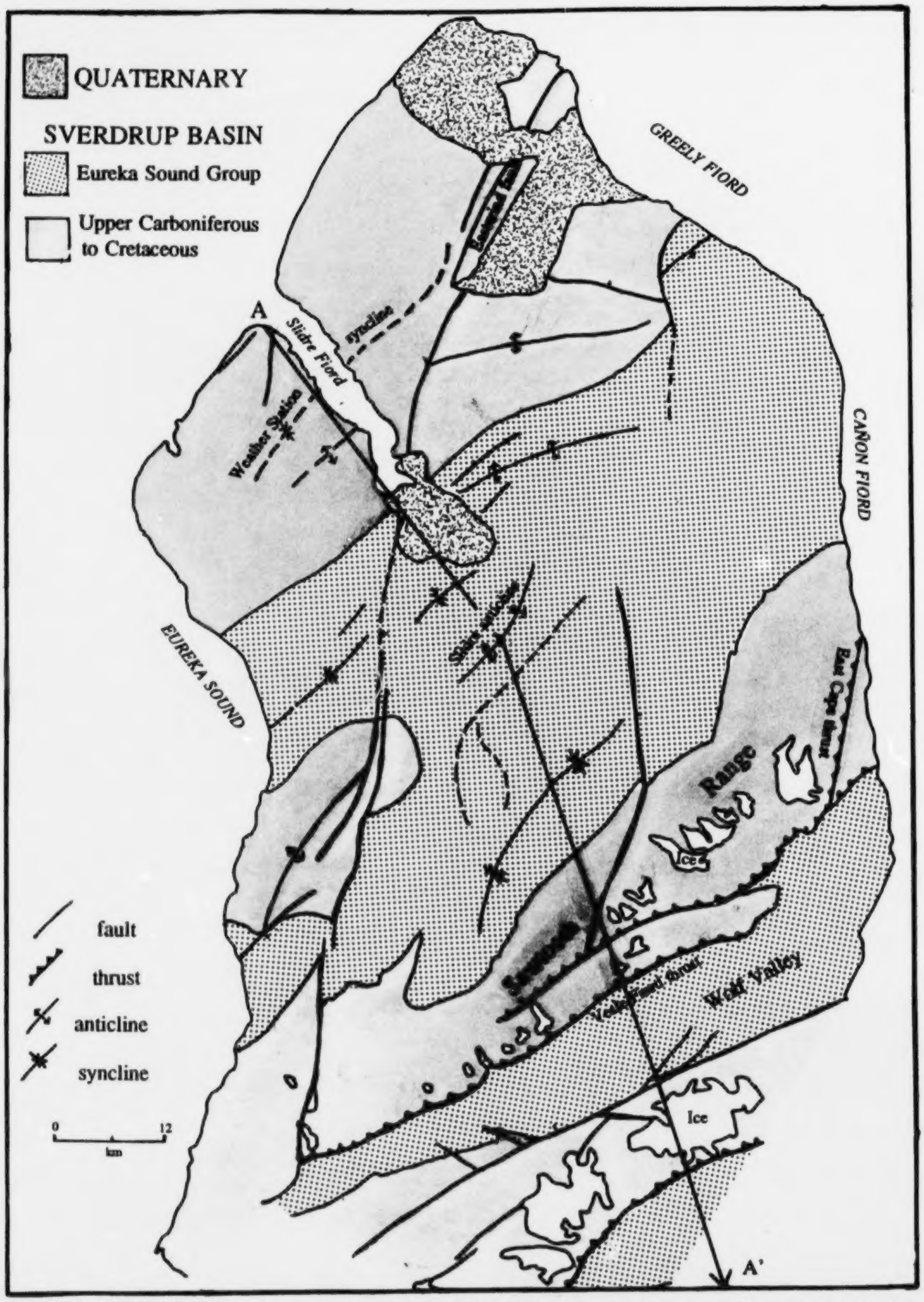

QUATERNARY
SVERDRUP BASIN
Eureka Sound Group
Upper Carboniferous to Cretaceous
GREELY FIORD
CAÑON FIORD
EUREKA SOUND
A
A'
Slidre Fiord
syncline
Weather Station
Slidre anticline
East Cape thrust
Sawtooth
Range
Vesle Fiord thrust
Wolf Valley
Ice
Ice
fault
thrust
anticline
syncline
0
12
km

Figure 2.6: Simplified geology map of western Fosheim Peninsula, showing general outcrop distribution and major structural features. Modified from GSC maps 1302A, 1308A, and 1311A.

Figure 2.7; Strike-aligned scarp-ridges on the western flank of Airfield ridge. Recessive shale and siltstone (dark coloured) form the troughs. Strata are steeply dipping due to evaporite doming.

Figure 2.8: Iceberg Bay Formation displaying sandstone-coal cycles. The Eureka Sound Group contains several billion tons of coal, some in seams more than 15 m thick (Ricketts and Embry 1986).

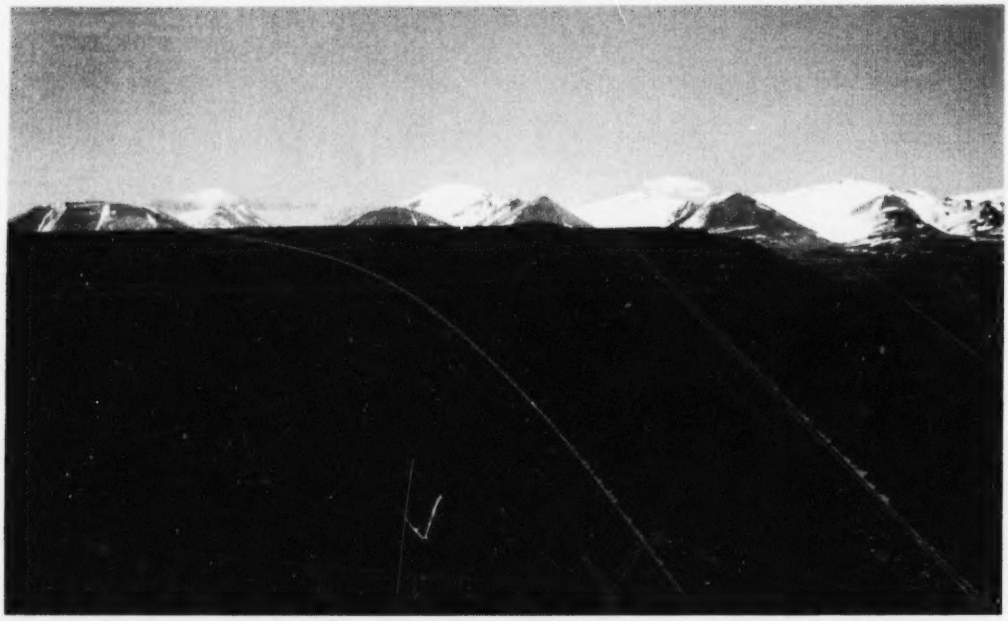


Figure 2.9: Erosional remnant of a former high plain that sloped northwestward from the Sawtooth Range (*background*) to ancestral Nansen Sound. Enclosed plant fossils suggest a Neogene age.

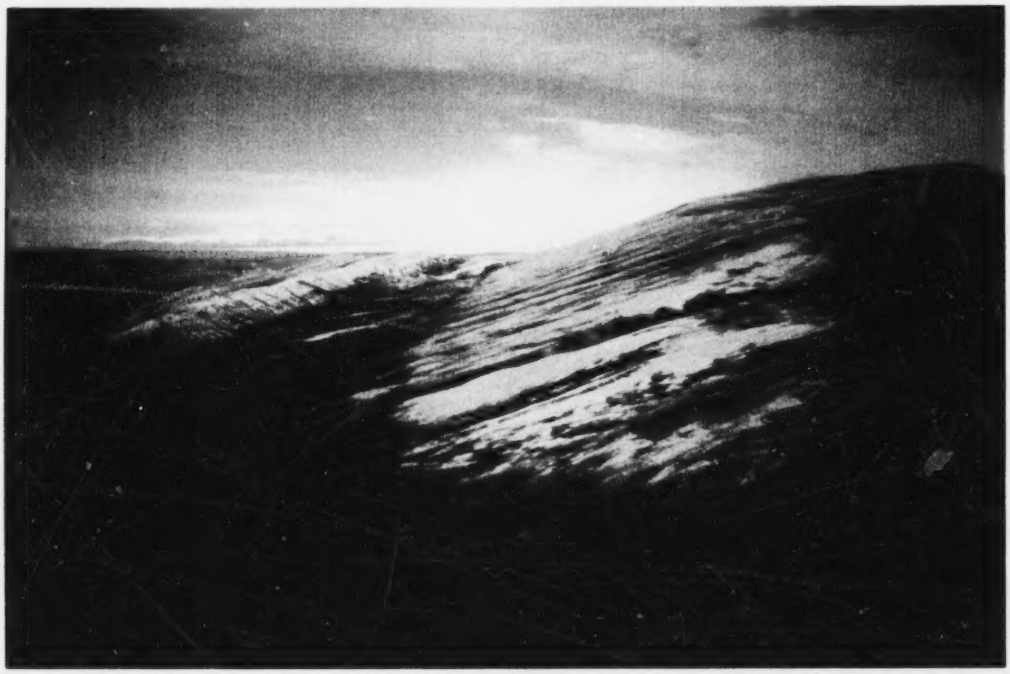


Figure 2.10: Thrust sheets in the Sawtooth Range were formed by compressive deformation during the climactic phase of the Eurekan Orogeny (mid Eocene).

prograded across west-central Ellesmere Island and eastern Axel Heiberg Island. Although these sediments have been loosely called Beaufort Formation (Tozer 1956; Thorsteinsson and Tozer 1970), they are referred to here as **high terrace sediments**, following Fyles (1990). These sediments consist of horizontally stratified, unconsolidated sand, silty sand and gravel lying beneath gently sloping gravelly terraces up to 600 m asl (Fig. 2.9; Fyles 1989). The terraces are remnants of a fluvial plain that sloped northwest towards Greely Fiord and Nansen Sound. High terrace sediments contain wood, peat, bone (beaver?) and other organic remains indicative of boreal forest to forest tundra conditions (Fyles 1989). Matthews and Ovenden (1990) used floral evidence to suggest that these deposits "span a great deal of time," perhaps from early Pliocene to early Quaternary. Thus deposition occurred after the Paleogene Eurekan Orogeny but before the incision by modern rivers. High terrace sediments on western Fosheim Peninsula are described in more detail in Chapter 4.

2.3 PHYSIOGRAPHY

Roots (1963) divided the physiography of Fosheim Peninsula into two zones: Ridges and Valleys of the Central Mountain Belt, and the Eureka Sound Uplands. The **Ridge and Valley** zone includes the Sawtooth Range and the mountainous region farther south where parallel to subparallel ridges are separated by straight valleys oriented NE-SW (Figs. 2.10 and 2.11). The relief there is >1000 m. Ridge crests are sharp to subrounded with an even crestline notched by gaps, gullies and cirques, somewhat like saw teeth (Fig. 2.12). Ridges and valleys are asymmetric in cross-section.

Drainage systems have either parallel or trellis networks with short tributaries. River gradients are low in the main valleys but it is apparent that significant incision has occurred because late Tertiary fluvial deposits are perched more than 300 m above the valley floors (Fig. 2.13).

The **Eureka Sound Uplands** is the second physiographic unit. It has a varied, subdued topography between 300 and 800 m asl and occupies the northern half of Fosheim Peninsula (Fig. 2.11). The broad summit of Airfield ridge, associated with Fosheim anticline (Figs. 2.6 and 2.11), was formed by evaporite doming near the base of Sverdrup Basin sediments (Fig. 2.5) and is likely an extension of diapiric ridges on Axel Heiberg Island. Black Top fault is part of a regional fault system that crosses the length of western Fosheim Peninsula and may extend northward beneath Greely Fiord to Blaa Mountain (Fig. 2.6; Geology Map 1311A). Black Top Ridge receives its name from the upthrown fault block which exposes dark coloured shale of the Blaa Mountain Formation (Fig. 2.14). Although generally recessive, this formation is given topographic expression through the strength of gabbro sills (*cf.* crest of Black Top Ridge (Fig. 2.14), along the spine of Mount Lockwood (Fig. 2.15)).

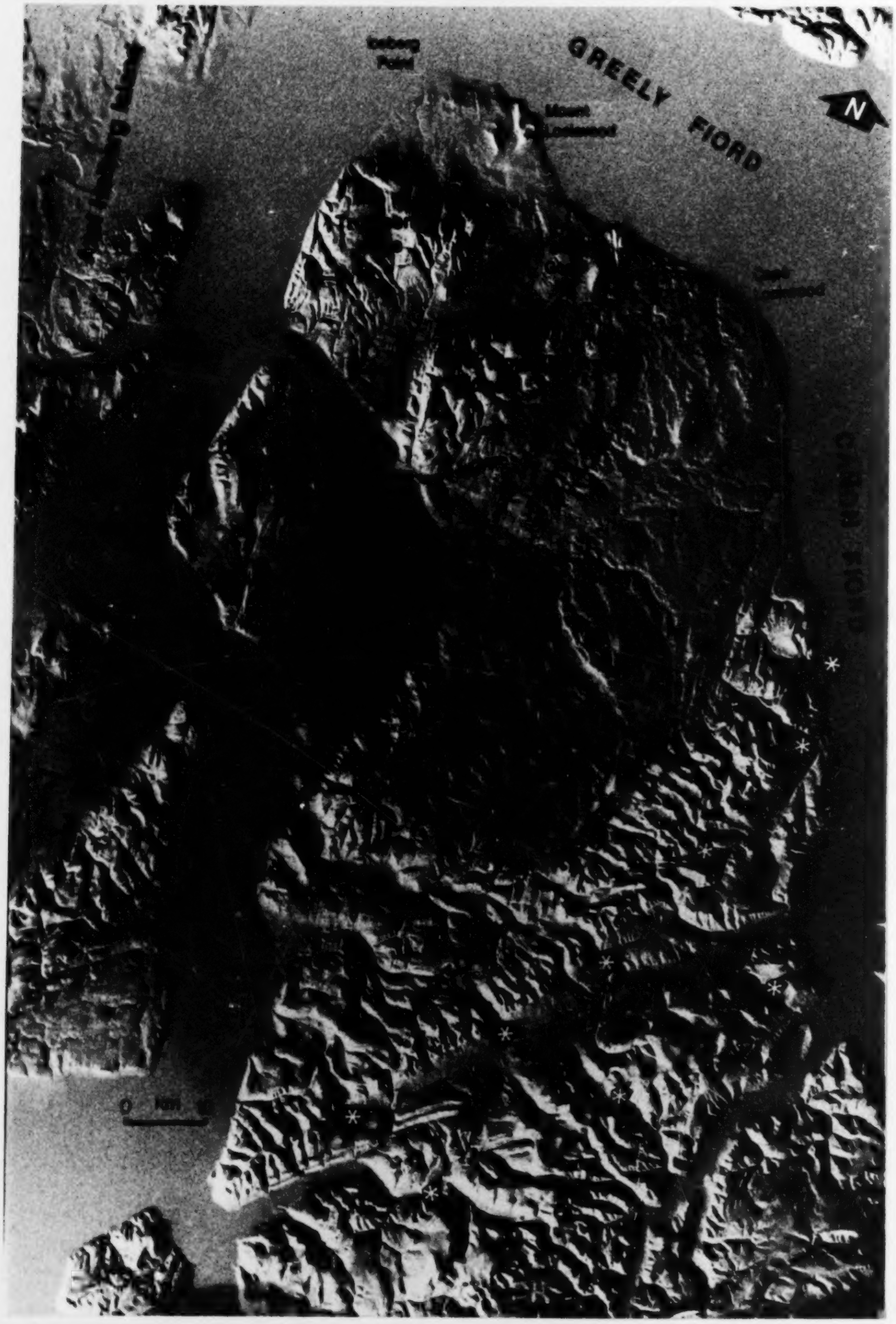

Iceberg Point
GREELY FIORD
N
Mount Lockwood
Cape Lockwood
CANON FIORD
0 km

Figure 2.11: Snow-covered, Landsat image of Fosheim Peninsula (April 1978). Physiographic features are indicated by official and unofficial placenames. The location and direction of late Tertiary drainage through the Sawtooth Range are outlined by arrows.

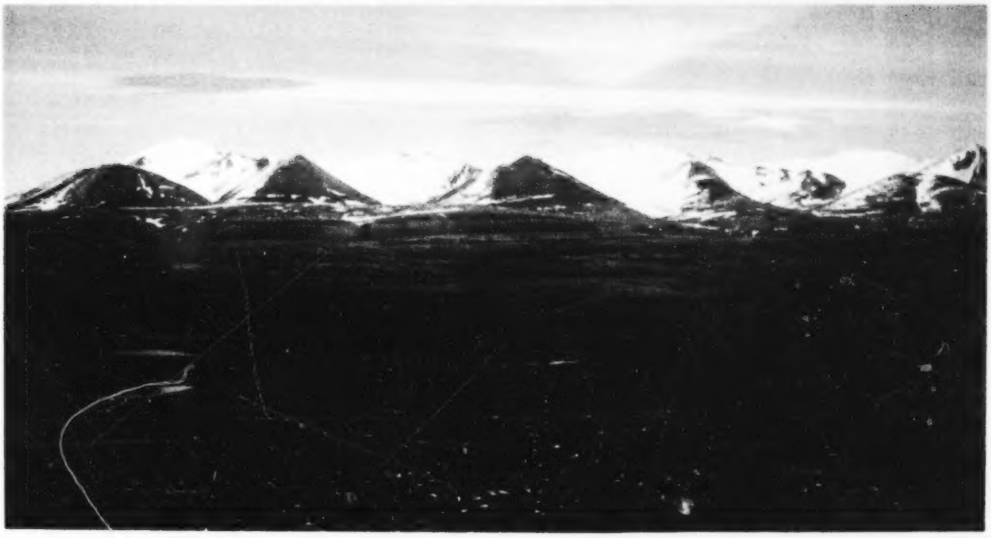


Figure 2.12: The ridge crest of the Sawtooth Range is notched by gaps giving the impression of saw teeth. Cirque basins contain the only glaciers in the study area.

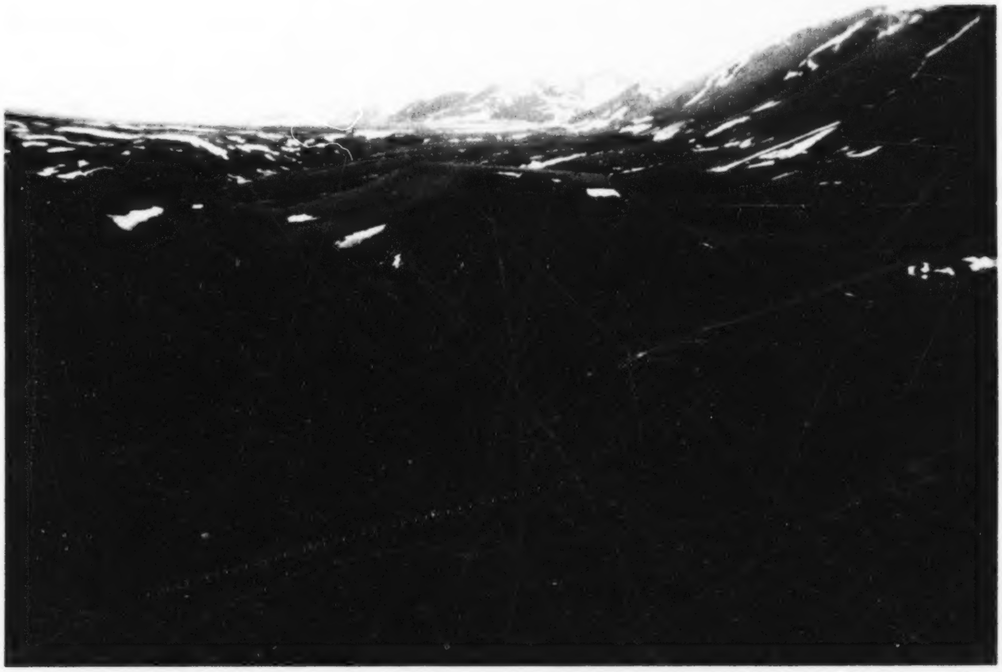


Figure 2.13: View southwest across a high terrace remnant in Wolf Valley. The terrace is preserved on a broad, flat plain at 450-500 m asl which may have been formed by rivers flowing across the valley (left to right) through gaps in the Sawtooth Range (right). Incision of this ancient plain and isolation of the high terrace >400 m above the modern valley was likely caused by subsequent drainage to a new base level, represented by Canon Fiord and Eureka Sound.

Figure 2.14: View looking NNE across Slidre Fiord towards Black Top Ridge. The ridge is an upthrown fault block, exposing shale and gabbro sills. Erosion of poorly consolidated sandstone and shale results in badland topography (*foreground*).

Figure 2.15: View north across till-blanketed Iceberg Point lowlands towards Mount Lockwood. It receives its topographic prominence from gabbro sills exposed by Eastwind fault.

The Eureka Sound Uplands generally exhibit a dendritic drainage system. Slidre River contains the largest catchment, including some glacierized basins in the Sawtooth Range. Many rivers have incised meanders in their lower reaches and, while some have reoccupied paleochannels, others have cut into bedrock during the Holocene. Gullying of poorly consolidated bedrock and raised marine fine-grained sediment has resulted in badland topography (Fig. 2.14).

2.4 CLIMATE

Fosheim Peninsula is considered the **"Garden Spot of the Arctic"** (Eureka weather station insignia). Recent vegetation and climate studies of the Queen Elizabeth Islands reinforce this view (Edlund and Alt 1989; Alt and Maxwell 1990). The peninsula's cherished reputation is a direct result of its topographic setting in the centre of an intermontane region (Fig. 2.1). The mountains that ring the Eureka Sound lowlands protect them from the incursion of cold Arctic Ocean air masses in summer and cyclonic activity in Baffin Bay. Broad-scale subsidence in the lee of the mountains causes low relative humidity and reduced cloudiness (Alt and Maxwell 1990) while the resulting rainshadow effect produces the lowest mean annual precipitation in Canada (Maxwell 1981).

The mean annual temperature range of 43°C at Eureka is the highest in the Canadian arctic (Atmospheric Environment Service 1982). Mean summer temperatures are anomalously high (5.4°C) for 80°N and emphasize the continentality of Eureka Sound. Furthermore, at Hot Weather Creek, central Fosheim Peninsula, the July temperature averaged 12.7°C during 1988 compared to 7.2°C at Eureka. The observed difference has been attributed to the increased continentality of Hot Weather Creek and the moderating effects of an open Slidre Fiord on Eureka (Edlund *et al.* 1989). The mean annual precipitation at Eureka is only 64 mm, 60-65% of which is in the form of snow. Most of the annual precipitation results from a relatively small number of synoptic events during early and late winter months (Bradley and Eischeid 1985).

Despite the severe aridity in the high arctic (Bovis and Barry 1974), the Eureka Sound intermontane region has the greatest density and diversity of vascular plants in the Queen Elizabeth Islands and rivals the vegetative productivity of southern Victoria Island at 70°N (Edlund and Alt 1989). This relatively lush vegetation reflects the higher summer temperatures, the longer snow-free period, and the higher melting degree day totals typical of the intermontane region (Alt and Maxwell 1990).

It is important to emphasize the strong regional gradients in the climate of the high arctic. For example, in contrast to Eureka Sound, northern Ellesmere and western Axel Heiberg islands are strongly influenced by the Arctic Ocean climate such that they experience the lowest summer temperatures, persistent fog and low cloud, and the shortest snow-free season. Consequently, these areas have the least diversity and lowest percentage cover of vascular plants and are dominated by herbaceous species (Edlund and Alt 1989). Such regional climate contrasts are not only reflected in vegetation patterns but also in the mass balance regimes of glaciers and ice caps. For instance, Meighen Ice Cap which borders the Arctic Ocean (Fig. 1.6) is characterized by low summer melt values and the occurrence of summer accumulation (Alt 1979), whereas outlet glaciers in Eureka Sound are subject to higher summer melt (Hattersley-Smith 1969). Likewise, the regional equilibrium line altitude (ELA) and glaciation level remain high over the intermontane region and clearly reflect the dominance of climatic variables over topography (*e.g.* July freezing level elevations; *cf.* Bradley 1975; Miller *et al.* 1975). Glaciation levels on Fosheim Peninsula and adjacent lowlands range between 800 and 1100 m asl.

2.5 PALEOCLIMATE

Paleoenviromental records from these distinct regions should also provide evidence of differing climatic conditions and responses. It is generally recognized that paleoglaciation levels were *ca.* 400 m lower than present across northern and central Ellesmere Island during the last glaciation (Hodgson 1985; England 1986; Lemmen 1989; Evans 1990; Sloan 1990). Considering the present pattern of glaciation levels, it is apparent that such a lowering would have least effect on the interior lowlands and most effect along the northwestern coasts of Ellesmere and Axel Heiberg islands. Geomorphic mapping in the lowlands surrounding Eureka Sound and Greely Fiord has demonstrated that large areas remained ice-free (Hodgson 1985; England 1990) whereas paleogeographic reconstructions on the north coast have documented almost total ice coverage during the last glacial maximum (see Fig. 11, Evans 1990).

Today, the predominantly ice-covered Arctic Ocean provides a limited moisture source in the Canadian high arctic; consequently, precipitation is the principal control on ice cap expansion (Bradley and England 1978; Miller and de Vernal 1992). Given that the climate was colder and drier during the last glaciation, then aridity was clearly the primary climatic constraint on the magnitude of the last ice advance in the high arctic (England 1987, 1990). Because the present precipitation pattern is topographically controlled, it is likely that a similar pattern prevailed during the last glaciation despite possible changes in circulation and synoptic conditions. Consequently, Eureka Sound would have received the least precipitation. It is not surprising that one of the largest recorded advances of a high arctic glacier during the last glaciation took place in Expedition Fiord, western Axel Heiberg Island (Fig. 2.1) where, among other factors, precipitation is 2-3 times higher (Lemmen *et al.* 1990).

The Eureka Sound intermontane area has the most favourable summer climate in the Queen Elizabeth Islands (Alt and Maxwell 1990). Here, the melt period begins the earliest, lasts the longest, and the ground is usually snow-free by late June. Because of this high spring melt intensity, it is suggested that glaciers and ice caps in the interior lowlands would experience the earliest and perhaps, the most rapid deglaciation in response to any increase in Northern Hemisphere solar radiation. Koerner and Fisher (1990), using ice core records from the Agassiz Ice Cap, documented a rapid change in $\delta^{18}O$ values at 10.75 ka BP and an abrupt increase in melting ≥9.5 ka BP. The earliest date on meltwater from an outlet glacier from the Agassiz Ice Cap is *ca.* 8.6 ka BP (England 1990).

2.6 LOCAL GLACIAL HISTORY

Most studies on the glacial history of west-central Ellesmere Island have concluded that the entire region was once inundated by ice. However, the timing of this glaciation has been disputed: Troelsen (1952) and Hattersley-Smith (1969) were undecided; Fyles (1962) and Boesch (1963) suggested an extensive pre-Wisconsinan glaciation; and Sim (1961), Roots (1963) and Hodgson (1985) discussed the possibility of an extensive last glaciation. Finally, England (1987) suggested that the most extensive glaciation predated the last glaciation and could have been as early as late Tertiary.

Details on the style of regional glaciation are sparse, usually confined to *isolated* observations on meltwater channels, striae, or erratics (Hodgson 1989). Assuming that these observations relate to the same event, then it appears likely that ice once flowed north along northern Eureka Sound striating bedrock (Fyles 1962; Rudberg 1963), and depositing granite erratics (Troelsen 1952; Sim 1961; Rudberg 1963) and fossiliferous diamicton (Sim 1961; Hodgson 1985). Hattersley-Smith (1969) described "a severe and widespread earlier glaciation" in central Ellesmere Island where a 1000 m thick outlet glacier advanced more than 20 km down Tanquary Fiord. In contrast, England (1987, p. 421) argued that "the lack of evidence for regional glaciation [in Greely Fiord], other than sparse erratics, suggests that this landscape is virtually nonglacial". He also argued against the complete inundation of Greely Fiord by any former trunk glacier.

Hodgson (1985, 1989), England (1990, 1992) and Sloan (1990) reported detailed studies on the style and extent of the last glaciation in west-central Ellesmere Island. Hodgson (1985) described abundant ice-marginal landforms in a 500 km long drift belt that lies 10-60 km west of the present Agassiz and Central Ellesmere ice cap margins (Fig. 1.7). This drift belt is generally confined to the fiord heads and is considered to be of paleoclimatic significance, representing the margins of expanded and coalesced ice caps between 9000 and 7000 BP.

England (1990, 1992) documented the limit of the last glaciation that interfingered with the sea around the perimeter of Greely and Cañon fiords (Fig. 1.9). Outlet glaciers from the Agassiz Ice Cap advanced only 5-60 km as they were constrained by severe aridity and by calving in fiords. The earliest recorded ice retreat was *ca.* 8000 BP whereas rapid deglaciation and ice unloading did not occur until 7200 BP.

Sloan (1990) described the inception of upland ice caps on southeastern Fosheim Peninsula during the last glaciation. Meltwater from these ice caps was draining into Cañon Fiord by 8000 BP. Sloan (1990) reconstructed possible deglacial and marine limits related to an extensive outlet glacier, flowing from Fosheim Peninsula into Cañon Fiord, predating the last glaciation. Glaciomarine sediments dated >29 000 BP were deposited by an ice shelf related to a former sea level between 147 and 164 m asl. This shoreline is 28 m higher than the Holocene marine limit and is one of the highest former sea levels recorded in the region.

2.7 RESEARCH OBJECTIVES AND RATIONALE

The objectives of this study can be divided into four broad categories, each of which includes specific questions arising from our current knowledge of the Quaternary history of adjacent regions.

(1) *TO DETERMINE THE EXTENT, CONFIGURATION AND AGE OF FORMER GLACIATIONS ON WESTERN FOSHEIM PENINSULA.*

Western Fosheim Peninsula lies within the zone of maximum thickness of the proposed *Innuitian Ice Sheet* which, according to Blake (1970), covered much of the Queen Elizabeth Islands during the last glaciation. It lies beyond the limit of the more restricted *Franklin Ice Complex*, an alternative model proposed by England (1976*a*) to describe the limited expansion of ice caps during the same interval. Although extensive fieldwork has firmly established the limit of the Franklin Ice Complex in surrounding areas, recent geophysical modelling of relative sea level data still favours a much larger ice load, equivalent to the Innuitian Ice Sheet, based on a deglacial history starting at 18 000 BP (Tushingham 1991). Western Fosheim Peninsula is critically located to test these models because it should possess either:

- recent evidence for complete inundation by the Innuitian Ice Sheet,

  or
- a deeply weathered landscape trimmed by the Innuitian Sea.

If fieldwork demonstrates that the peninsula was ice-free during the last glaciation, then it is likely that some evidence will be preserved of older, more extensive glaciations, as is the case on northeastern Ellesmere Island (England 1978). The limited information available on older glaciations of Ellesmere Island has inhibited any broad-scale reconstructions which might contribute to our understanding of high arctic paleoclimatology and landscape evolution.

(2) *TO DETERMINE THE MAGNITUDE, PATTERN AND AGE OF RELATIVE SEA LEVEL ADJUSTMENTS ASSOCIATED WITH FORMER ICE LOAD CHANGES.*

The magnitude of Holocene emergence (*ca.* 150 m) on west-central Ellesmere has been problematic in relation to the small ice load changes proposed during the last glaciation. On the other hand, the Innuitian Ice Sheet was originally proposed to explain, among other observations, a ridge of high emergence between Fosheim Peninsula and Bathurst Island (Blake 1970). England (1992) and England *et al.* (1991) have suggested ways in which the high emergence may be accommodated by a smaller ice load change. In this model, uplift from late deglaciation is maximized because it occurred after most of the global eustatic rise which, otherwise, would have diminished the emergence substantially. If western Fosheim Peninsula remained ice-free during the last glaciation, then the high emergence previously recorded there (150 m) cannot be explained through the unloading of a large ice sheet. Furthermore, there is a possibility that the transgressive history of the Innuitian Sea may be preserved somewhere on the peninsula.

The pattern of postglacial emergence has been used elsewhere to determine the configuration and retreat of former ice sheets (*cf.* Andrews 1970). Consequently, relative sea level data from western Fosheim Peninsula may contribute to the glacial and deglacial history of the area, assuming there was inundation by ice. Alternatively, the pattern of marine limit elevations may record a single shoreline that marks the limit of the full glacial Innuitian Sea occupying the peripheral depression beyond the last ice limit (England 1990, 1992). As such, isobases drawn on the limit of this full glacial sea may outline former centres of ice dispersal or other controls not yet fully understood (*cf.* England and Bednarski 1986). England (1992) described the isobase pattern on the Innuitian Sea in Greely and Cañon fiords as a narrow, plunging ridge which does not conform to the distribution of the last ice load and is unlike isobases drawn for other glaciated areas of arctic Canada. He proposed that the parallelism between the ridge crest and regional structural trends may indicate a neotectonic component in the emergence pattern, or that there was structurally-controlled, glacioisostatic unloading. Marine limit elevations from western Fosheim Peninsula will extend the database to Eureka Sound and, as a result, can test whether the ridge crest continues in this direction and whether it follows the structural trends.

The chronology and rate of emergence in west-central Ellesmere Island are not clearly understood. England (1992) presented a provisional emergence curve for Greely and Cañon fiords whereas Hodgson *et al.* (1991) constructed a minimum emergence curve for western Fosheim Peninsula. More control points are needed to resolve the relative sea level history and to provide increased resolution, particularly after 7000 BP. Dyke *et al.* (1991) demonstrated that good control on emergence curves can contribute to our understanding of postglacial rebound, particularly with regard to differentiating glacioisostatic and neotectonic components.

(3) *TO DETERMINE EARLY HOLOCENE GLACIER DYNAMICS IN THE CONTEXT OF GLACIAL GEOLOGIC, GEOMORPHIC AND PROXY PALEOCLIMATIC DATA*

Hodgson (1985) suggested that glaciers on west-central Ellesmere Island advanced to their last glacial maximum in the early Holocene (his model **B**). In contrast, England (1992) proposed that glaciers advanced to their maximum position between 14 000 and 9000 BP and that initial slow retreat commenced at 8000 BP. Both hypotheses have important implications regarding the response of high arctic glaciers to paleoenvironmental change. Because the early Holocene was the warmest period in the postglacial history of central Ellesmere Island (Koerner and Fisher 1990), and considering the increased warming predicted for the high arctic over the next century, it is critical to determine glacier responses under different paleoclimatic conditions and settings.

(4) *TO DETERMINE THE LATE TERTIARY AND QUATERNARY LANDSCAPE EVOLUTION OF WEST-CENTRAL ELLESMERE ISLAND*

There is considerable uncertainty regarding the evolution of the high arctic landscape since the deposition of the Eureka Sound Group in the mid Tertiary. Some geologists have argued that the high arctic has been tectonically inactive since the Eurekan Deformation (Miocene/Pliocene; Ricketts 1987*b*; De Paor *et al.* 1989), whereas England (1987, 1992) and Dyke *et al.* (1991) have presented evidence that contests this view. England (1987) argued that on a regional scale the Canadian arctic archipelago represents a Tertiary fluvial landscape developed on a contiguous landmass and subsequently fractured by block faulting; hence, the fiords are predominantly tectonic rather than glacial in origin. There is also the possibility that some of the postglacial uplift contains a neotectonic component. This study attempts to bridge the gap between mid-Tertiary and Quaternary paleoenvironments on Fosheim Peninsula by integrating the main geomorphic elements from this period into a general model of landscape evolution.

2.8 RESEARCH METHODS

The reconstruction of Tertiary and Quaternary paleoenvironments on western Fosheim Peninsula involved the investigation of glacial, marine and fluvial landforms and sediments. This included air photo interpretation, field mapping, and radiocarbon and amino acid analysis of organic samples. Air photo and field investigations provided information on surficial deposits and associated landforms which was later compiled on a surficial geology and geomorphology map. Surficial materials were described in a local and regional context. All stratigraphic sections observed were logged, interpreted and sampled using standard stratigraphic techniques, including unit thickness, structure, texture, and lithologic composition. Former ice margins, denoted by moraines, kames and meltwater channels, were mapped in relation to raised marine deposits. These marine deposits were investigated in terms of their stratigraphy, sedimentology, and organic content. Where possible, these sediments were related to former shorelines which were surveyed by altimetry. Mollusc shells and plant macrofossils collected from marine deposits were radiocarbon-dated to provide a chronology of late Quaternary glacial and sea level events. Ice-transported shells from till were analyzed for their amino acid content to develop a relative chronology of older events and to permit regional correlation.

2.9 DATING FORMER SEA LEVELS

In reconstructing the postglacial emergence history, it is important to make the correct interpretation of the paleosea level that pertains to dated samples. Shells from deltaic sediment normally relate to a former shoreline at the level of the delta terrace above the sample site, provided the terrace is not an erosional feature. Ideally, the sample should be obtained from Gilbert-style foresets which can be traced upwards to contemporaneous topsets; the point of intersection is commonly interpreted as the mean sea level at that time. In addition, shells from prodeltaic sediment also date the shoreline recorded by the delta terrace.

Shells obtained from nearshore sediments provide only minimum dates on former sea levels, unless they record deglaciation and rapid marine inundation of a depressed basin (overstep sequence, Dunbar and Rogers 1957; Bednarski 1988). In such a case, the oldest shell date relates to the establishment of the deglacial marine limit. Nearshore sediments in the peripheral depression beyond the glacial limit may show a coarsening-upward sequence in response to either **(1)** an advancing glacier, supplying increasingly coarse sediments; **(2)** a falling sea level; **(3)** a rising sea level with very high sediment input; or **(4)** a stable relative sea level where the rates of uplift and eustatic sea level rise balance (England 1992). In case (1) above, there should be independent evidence of ice advance in the form of disturbed sediments and redeposited shells. Case (2) should include evidence of marine offlap as the rate of sedimentation balances the rate of emergence, resulting in a normal seaward progression of terrigenous facies (Bednarski 1988). Case (3) may be recorded where local deglacial runoff preceded regional ice unloading and initial establishment of marine limit. Case (4) has been reported from several localities in Greely Fiord where shells predate the onset of observed emergence (England 1992).

2.9.1 Elevations of samples and paleoshorelines

Most elevations in the study area were determined by Wallace and Tiernan barometric altimeter, read to the nearest metre. As this instrument is sensitive to both temperature and atmospheric pressure changes, the elapsed time between readings was kept to a minimum and temperature corrections were made at each survey location.

In addition, each traverse was closed at a reference datum, either high tide line or base camp and, where possible, more than one reading was made over several days. The accuracy of this surveying technique is ± 2 m.

At several locations, elevation was derived from topographical maps in conjunction with air photographs. Given the map scale (1:50 000) and the contour interval (10 m), these elevations are considered accurate to ± 10 m.

Apart from potential measurement errors associated with site elevation, other error sources can affect paleosea level determinations. Of most importance are those related to the type of landform or sediment measured, and the relationship between their elevation and the contemporary sea level, including the tidal regime when they were formed (Andrews 1970; Rose 1978; Sutherland 1981; Gray 1983). Generally, the relationship between modern coastal landforms and present sea level is used as a guideline for the study of paleoshorelines; for example, Rose (1978) has shown that correction factors of up to 2 m must be applied to the elevation of specific landform types (*e.g.* beach ridges, washing limits) in order to reconstruct a former mean sea level. Moreover, the observational error associated with the point of measurement on the landform that records the evidence of a former sea level has been estimated to be ± 0.5 m (Labrador, Johnson 1985) and ± 0.3 m (Scotland, Sutherland 1981). Therefore, I have adopted Andrews' (1970) suggestion that a standard error of ± 1 m be applied to the corrected elevation of raised marine features to reduce variance and to achieve a conformable data set. As a result, an error margin of ± 3 m (which includes the instrument error) must be applied to marine limit elevations quoted in this study when former sea levels are being reconstructed.

2.9.2 Sample selection and preparation

The reliability of dating techniques is strongly dependent on the quality of material being used. Therefore, it is important to carefully select samples and to submit clean, uncontaminated material. Buried marine mollusc shells usually provide good samples whereas shells that have been lying on the surface for millennia usually have encrustations of secondary calcite and algae or lichen growth. These surface contaminants have younger ages than the shell material and must be mechanically removed. In addition, an acid leach removes at least the outer 20% of the shell material.

Clumps of detrital organic material were washed, sieved and examined under a microscope to select macrofossils for dating. The abundance of 'old', redeposited organics from Neogene terrace sediments in the study area necessitates careful selection.

2.9.3 Radiocarbon dating

Samples were radiocarbon-dated at the Geological Survey of Canada (GSC) using proportional gas counting, and at IsoTrace, University of Toronto, using atomic mass spectrometry (AMS). Radiocarbon is produced in the upper atmosphere by neutron bombardment of atmospheric nitrogen atoms. ^{14}C atoms are rapidly oxidized to $^{14}CO_2$ which mixes with atmospheric carbon dioxide and enters into all pathways of the biosphere. Over time, the ^{14}C radioactively decays to nitrogen, with the emission of a β particle. In living organisms, there is an equilibrium between their ^{14}C content and the atmosphere because there is a constant exchange of new CO_2. After death, however, this exchange ceases and the ^{14}C content declines at a negative exponential rate. The amount of ^{14}C remaining in a sample is indicative of its age.

In proportional gas counters, the carbon sample is converted into a gas which is then put into a 'proportional counter' capable of detecting β particles. In mass spectrometry, the actual number of ^{14}C atoms present are counted, not the very small numbers that radioactively decay during the measurement time (see extensive reviews by Bradley 1985; Linick *et al.* 1989; and Young 1991). The advantage of AMS dating is that it permits the precise analysis of very small samples, including individual valves and plant fragments.

GSC dates are normalized to 0‰ $\delta^{13}C$ for marine carbonate and to -25‰ $\delta^{13}C$ for wood and terrestrial organic material which builds in a 400 year correction for the marine reservoir effect. In some cases, GSC dates are 'uncorrected' (unnormalized) as $\delta^{13}C$ was not measured; however, differences between normalized and unnormalized GSC shell dates are almost always within the standard error of the age determination and are commonly only a few decades. Both sets of dates are comparable with regard to this study.

All IsoTrace dates have been corrected for fractionation to a base of -25‰ $\delta^{13}C$ but shell dates have also been adjusted for a reservoir effect of 410 years which is roughly equivalent to a fractionation correction to a base of 0‰ $\delta^{13}C$. Therefore, GSC and IsoTrace dates are directly comparable.

2.9.4 Amino acid analysis

Indigenous proteins and their constituent amino acids (bound within the carbonate matrix of fossil molluscan shells) are degraded in proportion to the length of time and amount of heat to which they have been subjected. This degradation includes the breaking down of high molecular weight proteins into smaller polypeptide chains, eventually releasing individual free amino acids (hydrolysis). For geochronological purposes, the most reliable reaction in protein diagenesis is racemization (or epimerization, in the case of the amino acid isoleucine). This reaction involves the inversion of amino acids from their protein L-configuration to their non-protein D-configuration. The ratio of the amino acid D-alloisoleucine to its diasteriomer L-isoleucine (**aIle/Ile**) measures the extent of epimerization in isoleucine. Epimerization rates depend on, among other factors, how amino acids are bound to each other, or if they are free (unbound). Rates are fastest when the amino acid is terminally bound and slowest when it is free. Consequently, the D:L ratios in the free fraction are higher than in the bound fraction. The alloisoleucine/isoleucine ratio in the total (free and bound) acid hydrolysate appears to be one of the best for geochronological applications (see reviews by Wehmiller 1984; Bradley 1985; and Miller and Brigham-Grette 1989).

To minimize variations caused by taxonomic differences in the rate of epimerization, data analysis in this study is restricted to the two most common molluscan genera used in aminostratigraphy: *Hiatella arctica* and *Mya truncata*. It has been demonstrated that both *Hiatella* and *Mya* epimerize at approximately equivalent rates within the range aIle/Ile = 0.1 to 0.5 (Kaufman 1992) and therefore, data from these two genera were pooled. Some shell samples comprising only fragments were identified to genus, based on their characteristic amino acid content.

Amino acid analysis was carried out by Dr. Julie Brigham-Grette at the Amino Acid Geochronology Laboratory (AGL), University of Massachusetts, Amherst. Results are displayed in Table C.2. Each ratio represents a mean value for at least three individual analysis on a valve or fragment. Normally, 3 to 5 valves or fragments were selected for analysis from each sample. International Laboratory Standards and AGL analytical standards are provided at the end of Table C.2.

CHAPTER THREE

Surficial geology and geomorphology

The surficial geology and geomorphology comprise the main source for reconstructing the Quaternary history of western Fosheim Peninsula. Surficial sediments are divided into six units based on their primary genesis. Spatial relationships between sediments and landforms are used to establish a relative sequence of Quaternary events while radiocarbon and amino acid analyses on organic samples provide additional chronological control.

The surficial geology map reproduced as Figure 3.1 (rear pocket) follows the general format of the Geological Survey of Canada (GSC) publications. It is based on sixteen weeks of field work during the summers of 1988, 1989 and 1990, stereoscopical mapping of air photographs (1:60 000, 1952), and unpublished data provided by D.A. Hodgson (GSC). Because of logistical constraints, only selected parts of the study area (5500 km^2) were visited. Initial fieldwork involved the classification of surficial units and the sampling of dateable material pertaining to glacial and marine sediments. Subsequent fieldwork focused on specific areas which were considered to be either critical to the reconstruction of regional Quaternary events, or representative of a larger landscape unit. As a result, large areas were mapped solely on the basis of air photo interpretation.

This chapter is divided into sections each of which describes a surficial unit and its associated geomorphology. Colour photographs are used in the text to illustrate sediment and landform characteristics whereas annotated air photographs provide examples of the spatial relationships between sediments and landforms (Appendix A). Stratigraphic sections are described and interpreted in Appendix B. For the purpose of description, the study area is subdivided into zones named after their principal physiographic feature (*e.g.* Hare Cape ridge, Iceberg Point lowlands). The significance of the surficial units to the interpretation of the Quaternary history is reserved for a later chapter.

3.1 WEATHERED BEDROCK/RESIDUUM

Weathered bedrock/Residuum is the most common surficial unit on western Fosheim Peninsula (Fig. 3.1). Bedrock is predominantly exposed on upland surfaces where gabbro sills form prominent ridges (Figs. 2.13 and 2.14), Mesozoic strata outcrop as thrust sheets (Fig. 2.11), or resistant sandstone caps recessive shale (Fig. 3.2). Residuum is preserved on low angled slopes and broad summits.

Weathered bedrock and residuum are widespread because the poorly indurated sandstone, siltstone and shale are strongly susceptible to weathering. The Eureka Sound Group is the most commonly exposed bedrock and consists predominantly of poorly cemented, quartzose sandstone interstratified with shale, siltstone, mudstone and coal. Outcrops are generally mantled by a medium-grained sand to silt-size residuum. Mesozoic sandstone-dominated formations are commonly overlain by extensive areas of rubble, composed of sandstone blocks and fine- to medium-grained sand. More recessive, shale-dominated formations weather to platy or crumbly fragments, silt and clay.

Residuum is reported to be 1-2 m thick in various parts of western Fosheim Peninsula (Hodgson *et al.* 1991) and although its texture reflects the lithology of the underlying rock formation, there is some surface reworking and mixing of grain sizes. Erratics lie scattered on the surface of bedrock and residuum in all parts of the study area and in places, they also show advanced weathering like the bedrock. Below marine limit, bedrock surfaces have been washed and local residuum reworked with littoral and nearshore sediment. Eolian activity has also modified residuum through the deflation of fine grained sediment as evidenced by surface lags and incipient stone pavement (Fig. 3.3). Ice wedge development and the formation of polygonal nets have altered residual surfaces, especially in fine-grained material. Slope processes have produced sorted and non-sorted stripes in residuum, and gelifluction lobes commonly incorporate residual material.

Although weathering processes are poorly understood in cold, arid environments, three dominant mechanisms have been proposed. **Frost wedging** involves the mechanical breakdown of rocks by freezing of water within pore spaces, joints, and bedding planes (French 1976). Important controls on the effectiveness of frost wedging are moisture availability and rock lithology. Although western Fosheim Peninsula is classified as a polar desert (Bovis and Barry 1974; Bliss and Bliss 1984), this does not provide an accurate impression of surface moisture conditions. Wind blown snow is swept from exposed ridge crests and preferentially accumulates on lee-side slopes and in valleys where snowbanks may persist for most of the year (Woo *et al.* 1990, 1991). Meltwater and summer rainfall are confined to the active layer (10-80 cm thick) by permafrost and consequently, lower slopes and valley bottoms generally receive moisture during critical freeze-thaw periods. Lithological factors

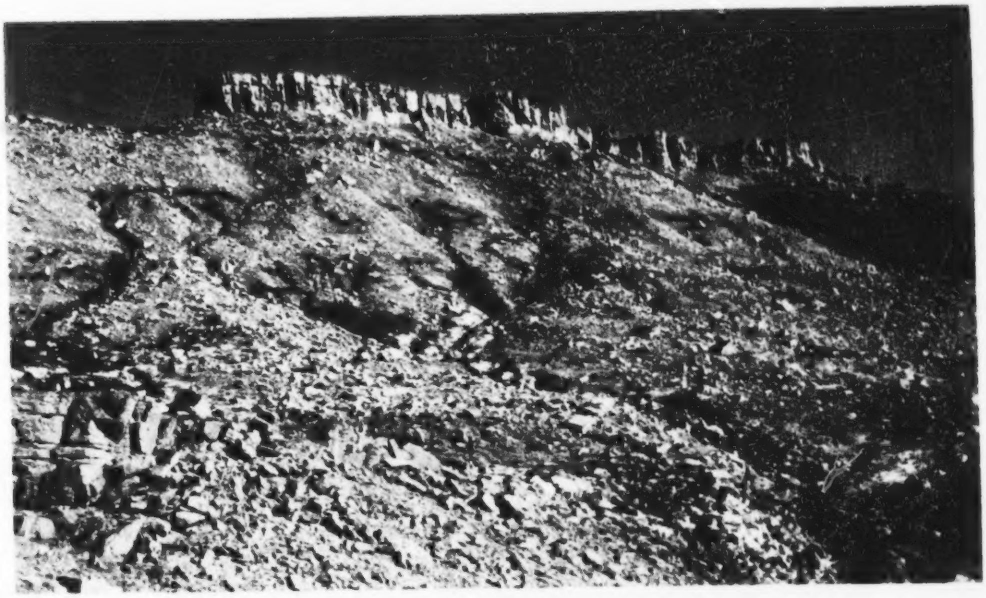


Figure 3.2: Resistant sandstone ridge capping recessive shale, south of Slidre Fiord.

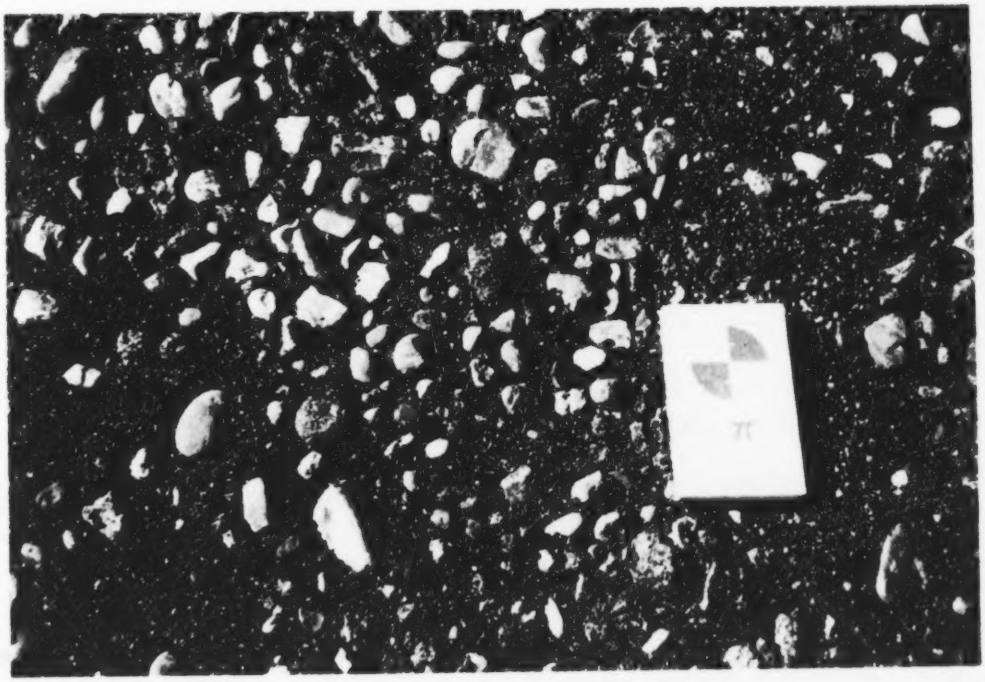


Figure 3.3: Incipient stone pavement composed of erratics in residuum near Hare Cape Ridge.

such as planes of weakness and grain size distribution are also important in frost wedging (Potts 1970).

Hudec (1973) argued that **hydration shattering**, involving the adsorption and desorption of pore water, may be the dominant mechanism in rock breakdown in arctic environments. This process does not require the freezing of pore water nor extremely low temperatures and operates most effectively on fine-grained, argillaceous rocks. The occurrence of shale felsenmeer and extensive scree slopes below shale and siltstone outcrops on western Fosheim Peninsula suggests that both frost wedging and hydration shattering may be active although their relative importance cannot be gauged at present.

Fahey (1985) has demonstrated that **salt weathering** is effective in extremely cold environments. Salt crystal growth and the hydration and rehydration of anhydrous salts cause rock splitting and disintegration. Sandstone and limestone are apparently more susceptible to salt attack than shale because of their water absorption capacity and microporosity (Cooke 1979). Wellman and Wilson (1965) and W.C. Bradley *et al.* (1978) emphasized the role of salt weathering in the development of tafoni (cavernous weathering) in the dry valleys of Antarctica and a semi-arid region of southern Australia, respectively. Watts (1983) attributed the formation of pits, flakes and tafoni to salt weathering in granite on southeastern Ellesmere Island. Similar features have been observed on western Fosheim Peninsula and are likely the result of salt weathering (Fig. 3.4).

Little is known about weathering rates in cold, arid environments and therefore, it is difficult to assess the age of the bedrock weathering features on western Fosheim Peninsula. In addition, rates vary with changes in climate. For example, the depth of weathered bedrock recorded from drill sites on western Fosheim Peninsula exceeds the present depth of the active layer, likely indicating that this deeper weathering formed during a warmer period when the active layer was deeper (Hodgson 1989).

Criteria such as weathering rind thickness, macro and micro pit size, surface relief, silica to aluminum weathering ratios, and numbers of intensely weathered erratics have been used to suggest a lengthy period of subaerial weathering of the granitic uplands in southeastern Baffin and Ellesmere islands (Andrews and Miller 1972; England and Andrews 1973; Boyer and Pheasant 1974; England 1976*a*; England *et al.* 1978, 1981; Dyke 1979; Watts 1983). Sugden and Watts (1977) and Watts (1983) proposed that these weathering features may have survived the last glaciation, protected by non-erosive, cold-based ice. Although the features described here for western Fosheim Peninsula are similar to those reported by Watts (1983), they occur in soft sedimentary bedrock, not granite. Consequently, the concept of a protective, cold-based ice sheet is not likely applicable to the sedimentary terrain of the study area (S.H. Watts, pers. comm. 1988). Furthermore, Dyke (1983) has shown that cold-based ice is quite erosive leaving widespread evidence in the form of ice-

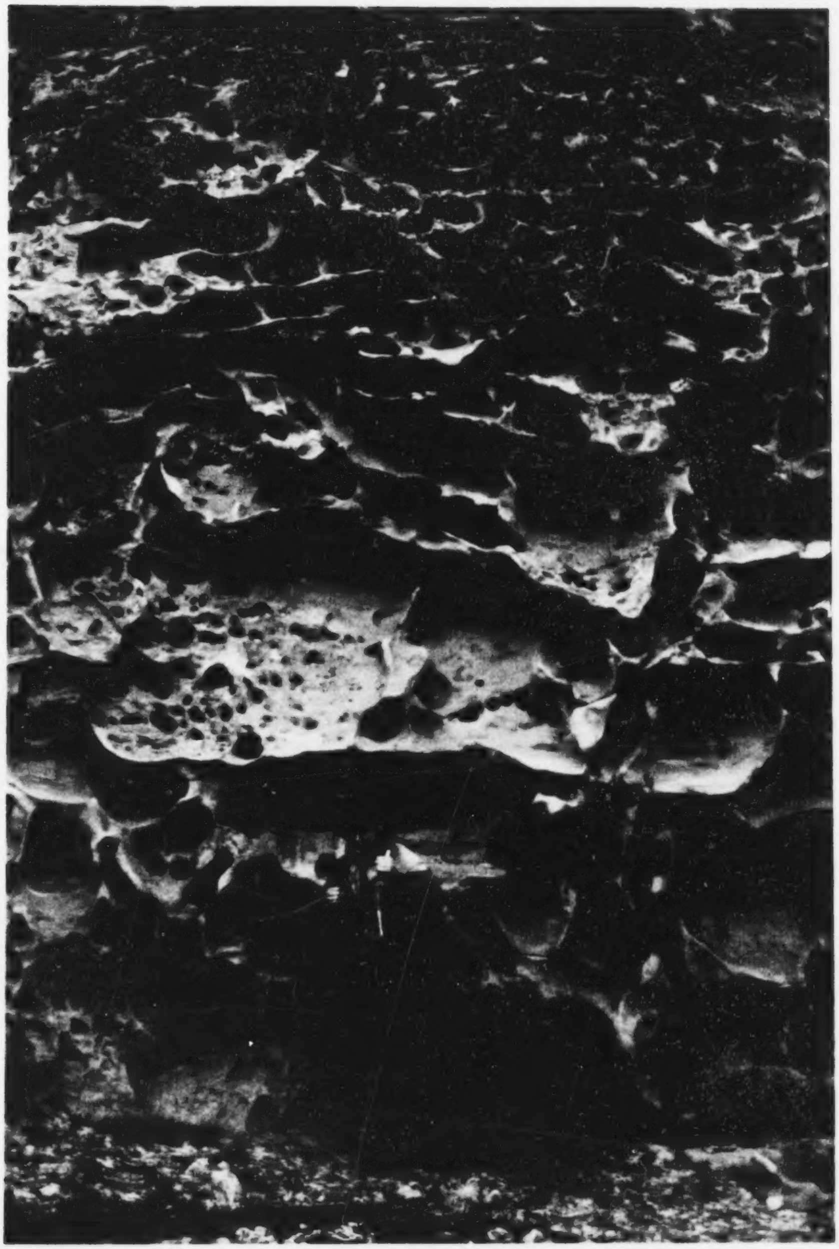


Figure 3.4: Cavernous weathering (tafoni) in coarse-grained, calcareous sandstone near Hare Cape ridge. Salt crystallization is the probable weathering mechanism.

marginal meltwater channels. Therefore, the occurrence of lowland tors (Fig. 3.5), intensely weathered bedrock (Figs. 3.4 & 3.6), and rooted, coalified tree stumps projecting more than 1 m from the ground (*cf.* McMillan 1963) suggest that there has been considerable dissection and a substantial period of subaerial weathering (that is, much longer than the Holocene) since the last extensive glaciation of western Fosheim Peninsula.

3.1.1 Associated landforms

Considering the nature of the soft sedimentary bedrock on western Fosheim Peninsula, it is unlikely that evidence of former glaciation is preserved except on more resistant rocks. Rare striations recorded by Fyles (1962) on gabbro sills near Blue Man Cape indicate a northward flow of ice along northern Eureka Sound. In this study, striations and grooves were observed only on a resistant sandstone outcrop near the northern end of Black Top Ridge where they outlined a northwesterly ice flow.

Meltwater channels are also preserved only on resistant bedrock (Fig. 3.1). At the northern end of Black Top Ridge, ice-marginal channels have been eroded into Heiberg Formation sandstone (Figs. A.1 and A.2). These V-shaped channels are up to 50 m deep, 250 m wide and in some cases can be traced for 5 km, descending from 600 m asl to the adjacent lowlands (170-180 m asl). They display an intersecting pattern with younger, diagonal channels superimposed on older, subhorizontal ones. Together with corresponding channels on Snow Goose hill, they record the systematic eastward and southeastward retreat of a regional ice cover. Above Eastwind Lake, meltwater channels occur up to 280 m asl on the western slope of Black Top Ridge marking the retreat of ice from the lake basin onto Weather Station lowlands (Fig. 3.1).

On the northeastern end of the peninsula, a series of subparallel meltwater channels lie along an escarpment at *ca.* 300 m asl and descend northwestward to Hot Weather Creek and Greely Fiord (Figs. 3.1 and A.3). Although these channels are eroded primarily in bedrock, they also incise high terrace remnants and till. At higher elevations, the channels form wide, steep-sided valleys whereas lower down they become more shallow and appear as overfit valleys. Because the channels originate at the scarp edge, a former ice margin must have abutted the scarp producing meltwater that eroded the channels.

Channels originating at the base of the escarpment descend gently towards Slidre River valley and Cañon Fiord where marine limit deltas occur (Figs. 3.1 and A.3). Likewise, the larger meltwater channels have accumulations of fine-grained sediment in their lower reaches, marking marine limit. This suggests that an ice cap existed on the upland surface above the escarpment during the last glaciation. During

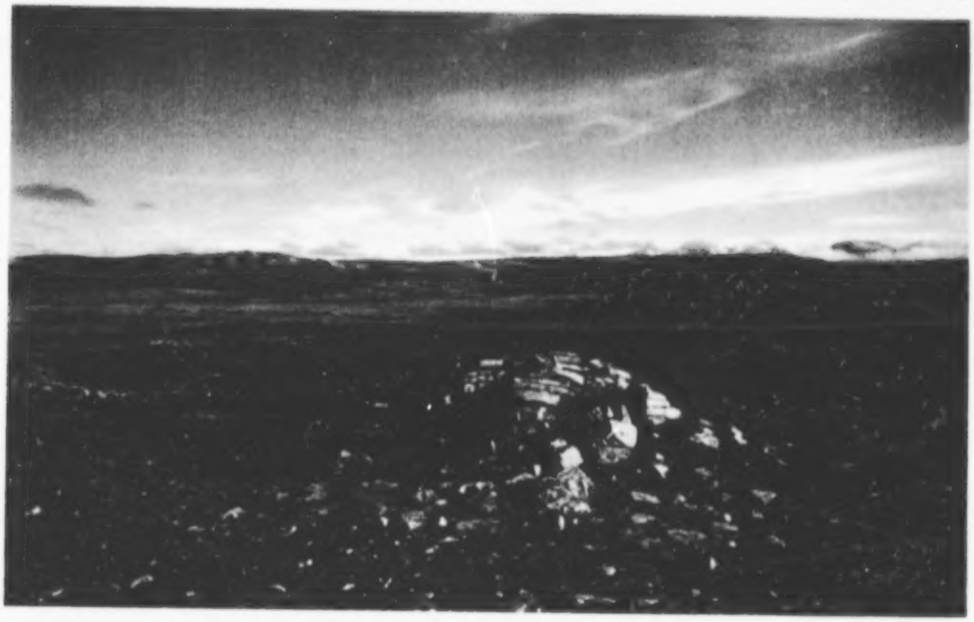


Figure 3.5: Sandstone tor at 183 m asl in the central lowlands of western Fosheim Peninsula.

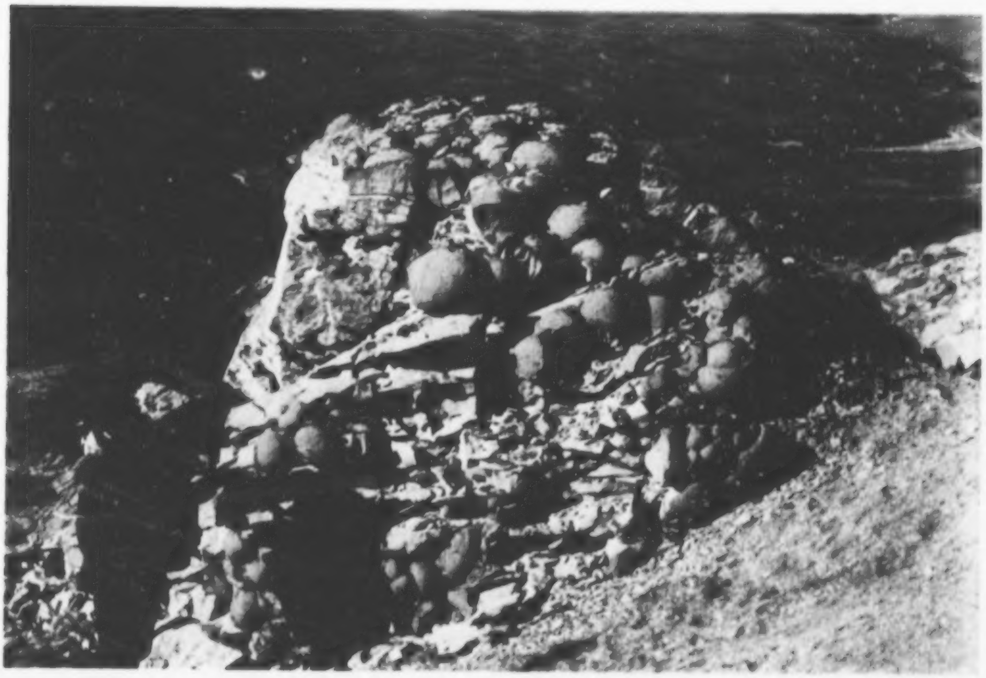


Figure 3.6: Nodular weathering on sandstone which appears to be a near-surface phenomenon. Note recent desiccation cracking of nodules.

deglaciation, northwestward-draining meltwater reoccupied the older meltwater channels whereas meltwater flowing off the escarpment eroded shallow channels on the till-covered lowlands.

Elsewhere, meltwater channels were eroded in bedrock or reoccupied during the demise of former upland ice caps on local ridge summits (Fig. 3.1). Many of these channels now serve as overfit stream valleys but large boulders strewn on channel floors attest to higher meltwater discharges. Nested meltwater channels along the foothills of the Sawtooth Range record the retreat of former ice that drained through existing valley systems. Ice-marginal channels in a low pass between Mount Bridgman and the adjacent unnamed peak indicate the eastward retreat of an ice lobe into Cañon Fiord (Figs. 3.1 and A.3).

3.2 HIGH TERRACE SEDIMENTS

Fyles (1962) first reported unconsolidated deposits beneath high terraces on west-central Ellesmere Island and later, on the basis of plant macrofossils, deduced that they were of Neogene age (Fyles 1989). High terraces have been mapped on eastern Axel Heiberg Island and west-central Ellesmere Island (Hodgson 1974; Hodgson and Edlund 1977). On western Fosheim Peninsula, remnants are preserved on the central lowlands and along the foothills of the Sawtooth Range (Figs. 3.1 and A.3). Although high terrace sediments rarely outcrop, they are nevertheless treated as a surficial unit because they form conspicuous mesas and buttes and rim prominent plateaus (Figs. 2.17, 3.7 and 3.8). These landforms are more resistant to erosion because terrace sediments are better drained than the underlying bedrock and a veneer of gravelly diamicton armours terrace treads and risers (Hodgson *et al.* 1991).

3.2.1 Nature and Occurrence

Terrace sediments commonly occur as either isolated remnants on bedrock ridges or dissected rims of prominent plateaus. Terraces have flat treads and steep risers but are rarely continuous for more than several hundred metres. They range in elevation from 360 to 150 m asl; however, this does not reflect a regional slope because observations were made on erosional remnants. Secondary terraces eroded into the sides of some remnants are interpreted as glaciofluvial in origin (Fig. 2.17). In other cases, glacial meltwater has dissected terrace remnants producing flat-bottomed, steep-sided channels which now contain shallow ponds (Fig. 3.9).

Some till-covered plateaus exhibit a dissected rim of terrace sediments (Fig. 3.8). It is unknown to what extent terrace sediments underlie the till surface but in some cases, bedrock is clearly visible through till veneer. Modern streams flowing off the plateaus dissect terraces and underlying bedrock. In one of these tributary valleys

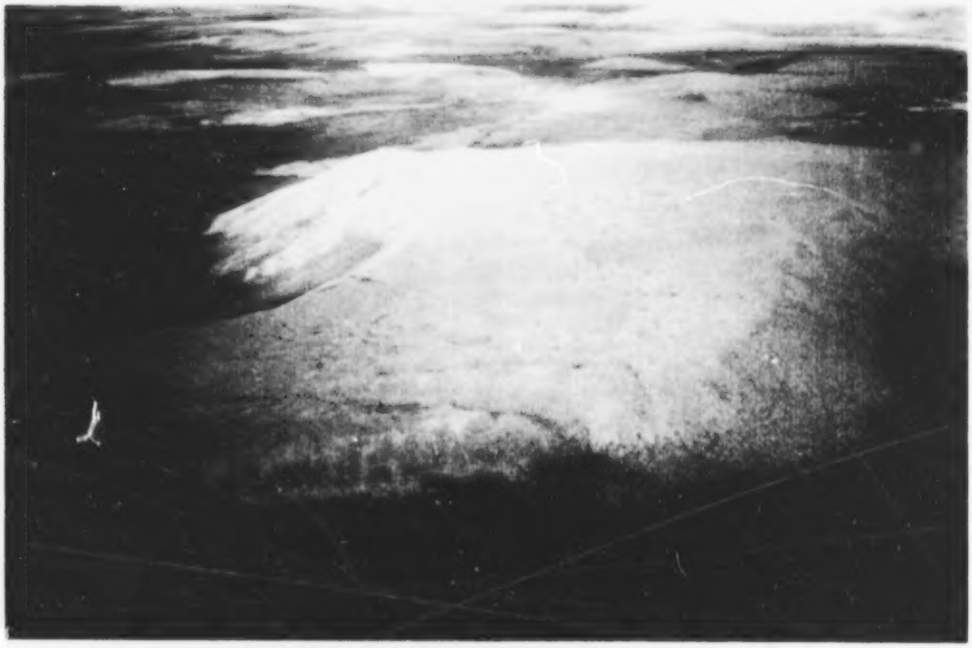


Figure 3.7: Isolated remnant of high terrace sediments armoured by gravelly diamicton. Its steep slopes are slowly retreating due to slumping. Gelifluction is active on the footslope.

Figure 3.8: Edge of till-covered plateau at 180 m asl exposes high terraces sediments overlain by gravelly diamicton. The diamicton is considered to be the coarser equivalent of the dark toned, vegetated till surface on the centre of the plateau (bachground).

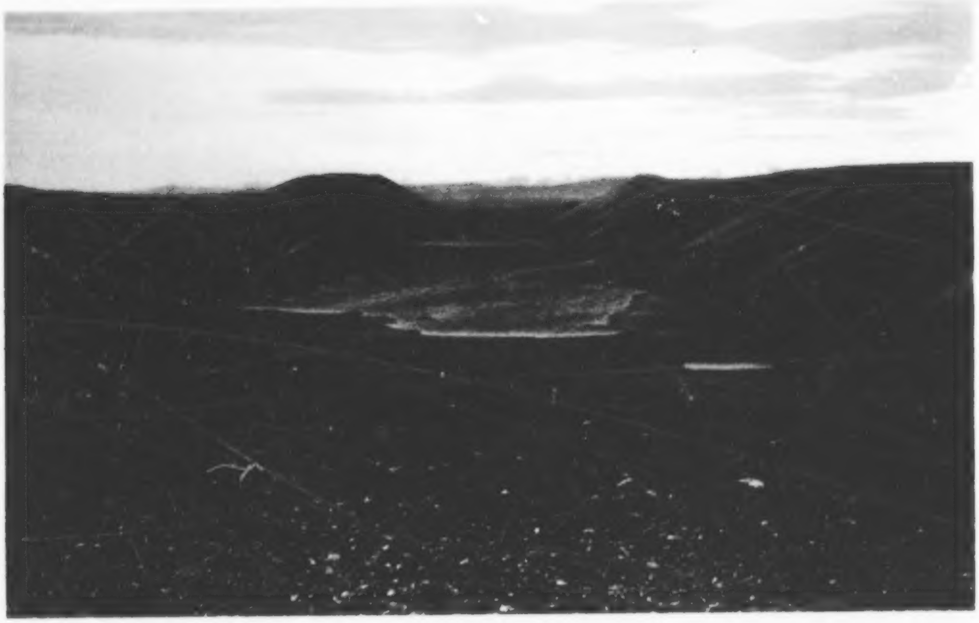


Figure 3.9: Meltwater channel eroded in high terrace sediments at 330 m asl. Unaltered wood was recovered from slumped material near this site.

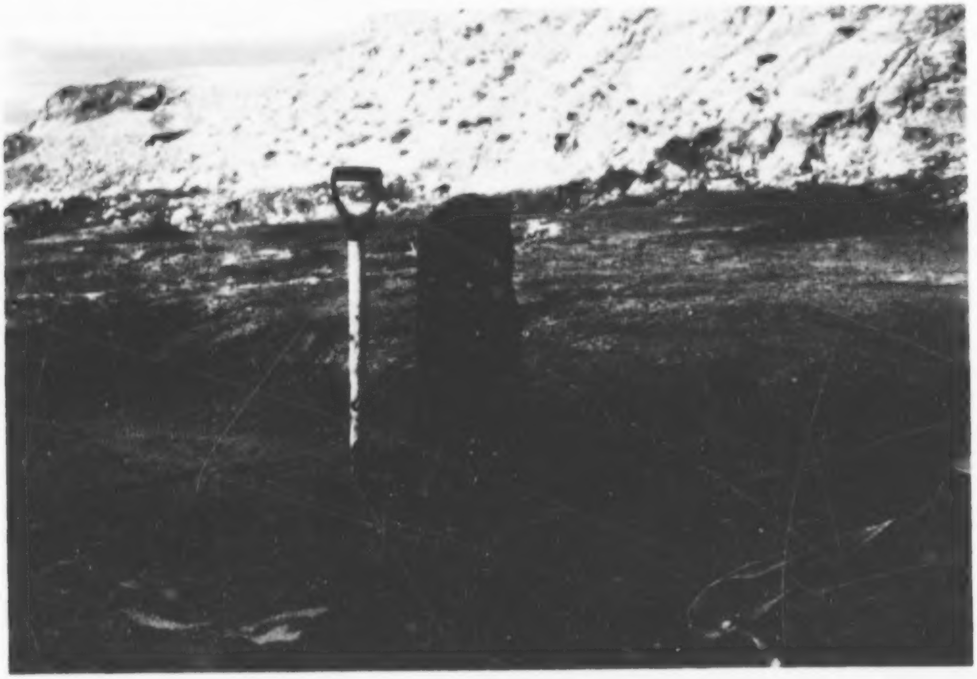


Figure 3.11: Examples of unaltered wood associated with or recycled from high terrace sediments on western Fosheim Peninsula. Shovel is *ca.* 1 m long.

between Eureka Sound and the Sawtooth Range, *in situ* tree trunks are preserved in fine-grained terrace sediments (R. Marris, pers. comm. 1989).

Where this topographic distinction is less obvious and sections are absent, remnants of high terrace sediment overlain by gravelly diamicton are indistinguishable from glaciofluvial landforms composed of gravelly outwash. On northeastern Fosheim Peninsula, for example, mesas and flat-topped terraces appear to form a lower, dissected plain (elevations range from *ca.* 250 to 175 m asl) on the coastal lowlands (Fig. A.4). Although these landforms are kame-like and are overlain by a poorly sorted gravel lag, including large angular blocks, there was no opportunity to determine whether they were gravelly outwash or fine-grained, Neogene terraces. However, it is possible that these lower terraces are composed of sediment recycled from higher terraces by fluvial or glaciofluvial processes during the Neogene or Quaternary. Their abrupt termination at or above 175 m asl may suggest that they graded to a former sea level at this elevation; therefore, the lower terraces may represent a dissected delta surface. P.F. Bruggeman collected wood from the top of a gravel ridge in this general vicinity (Broecker *et al.* 1956). According to the site description, it must have been above Holocene marine limit on one of the gravel terraces. The wood, which Bruggeman described as driftwood, was dated >33 000 BP (L-266C; Broecker *et al.* 1956). Although the origin and age of this sample suggests a pre-Holocene high sea level, there is a possibility that the wood was redeposited from Neogene terraces. For the moment, all the gravel terraces are arbitrarily portrayed on the surficial geology map as *high terrace sediments* but require further investigation to resolve a more detailed classification.

Although exposures of terrace sediments are rare, Hodgson *et al.* (1991) reported a 27 m thick section in a deep gully near Hot Weather Creek. The exposure contained "generally planar bedded, in places crossbedded, fine sand, subordinate sandy silt, granule beds and coal fragments" overlain by a veneer of bouldery gravel. High terrace sediments elsewhere on west-central Ellesmere Island have a similar fine-grained texture (Fyles 1989).

Near Cape Lockwood on the northern coast of Fosheim Peninsula, a deeply incised river valley has exposed 70 m of gently dipping beds of greyish sand overlain by 1-2 m of gravelly diamicton with striated erratics (section 1, Appendix B). The sand is tentatively interpreted as glaciofluvial in origin. Its uniform texture implies a steady sediment supply which was recycled from Neogene terraces. The presence of unaltered wood in the sand supports this conclusion (Fig. 3.11). The gravelly diamicton capping the section is interpreted to be till. Its occurrence precludes a Holocene age for the sandar and implies that the incision of Neogene terraces was underway prior to the last regional glaciation of the peninsula.

3.3 TILL

The regional till sheet on western Fosheim Peninsula is divided into two mappable units: blanket and veneer (Fig. 3.1). Till blanket is sufficiently thick (>2 m) to completely obscure the underlying bedrock structure (Fig. 3.12), whereas the surface of till veneer mimics the underlying rock surface (Fig. 3.13; Dyke 1983). A discontinuous till veneer describes scattered erratics resting on bedrock and residuum (Figs. 3.14). Although it does not form a mappable unit, erratics were observed on all bedrock and residuum surfaces in the study area.

Regionally extensive blankets and veneers of stony diamicton are interpreted as till because: **(1)** the base of the diamicton commonly forms an abrupt, planar contact with the underlying sediment and is considered to be erosional; **(2)** sub-diamicton sediments are penetratively deformed consistent with glacial shearing and loading; **(3)** diamictons contain incorporated clasts and/or rafts of underlying sediment; and **(4)** diamictons contain striated, regional erratics and in some cases, transported shell fragments.

3.3.1 Provenance

Till contains lithologies that may have either a regional or local provenance. Granite and quartzite erratics have their nearest source in the northernmost Canadian Shield on Ellesmere Island, 120 km southeast of the study area. Limestone and dolomite erratics originate in Permo-Carboniferous strata whereas chert pebble conglomerate can be traced to the Carboniferous Canyon Fiord Formation (H. Trettin, pers. comm. 1991). All of these are exposed along thrust faults within 20 km south or east of the study area. Diabase is a common erratic originating locally from diabase dykes and sills which intrude Mesozoic rocks on west-central Ellesmere Island. Large sandstone blocks were transported short distances from the locally extensive Triassic Heiberg Formation. Limestone, dolomite, sandstone, diabase, and chert pebble conglomerate are common erratics in till but show no distinctive distribution on western Fosheim Peninsula.

Quartzite erratics are present on all till surfaces in the study area and are typically subrounded, displaying crescentic scars like percussion or chatter marks (Fig. 3.15). Because glacial modification (*e.g.* striations, facetting) appears minimal, some quartzite erratics may have a fluvial origin with only minor reworking by ice. A possible source for these erratics is the high terrace sediments which originated in the mountains of east-central Ellesmere Island. Although only sandy deposits are associated with terrace remnants on western Fosheim Peninsula, pebble to cobble gravel overlie sand and silty sand above Strathcona Fiord (Fyles 1989). If equivalent gravel beds were eroded and redeposited as gravelly diamicton on western Fosheim

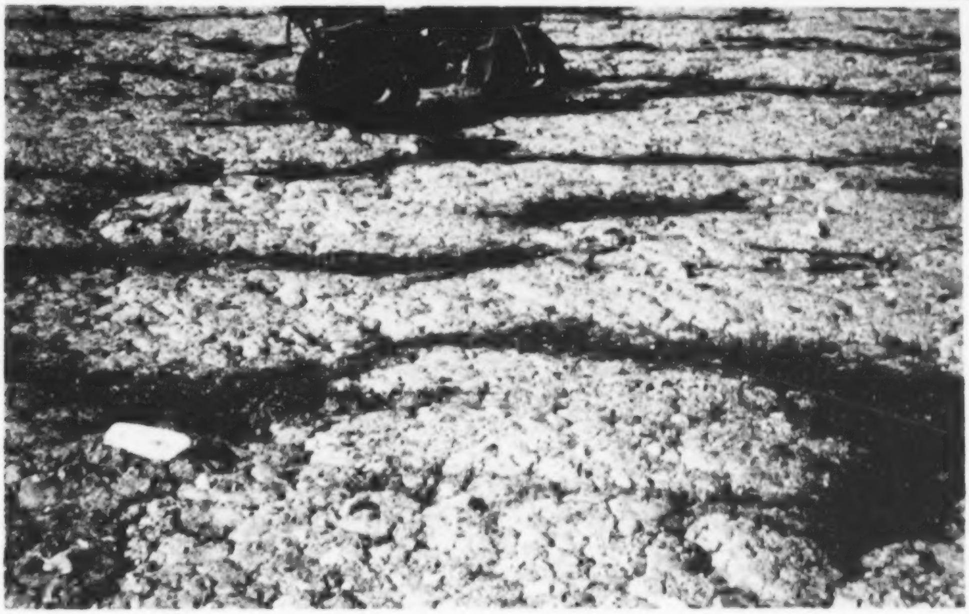


Figure 3.12: Till blanket obscuring underlying bedrock structure on the central lowlands (276 m asl).

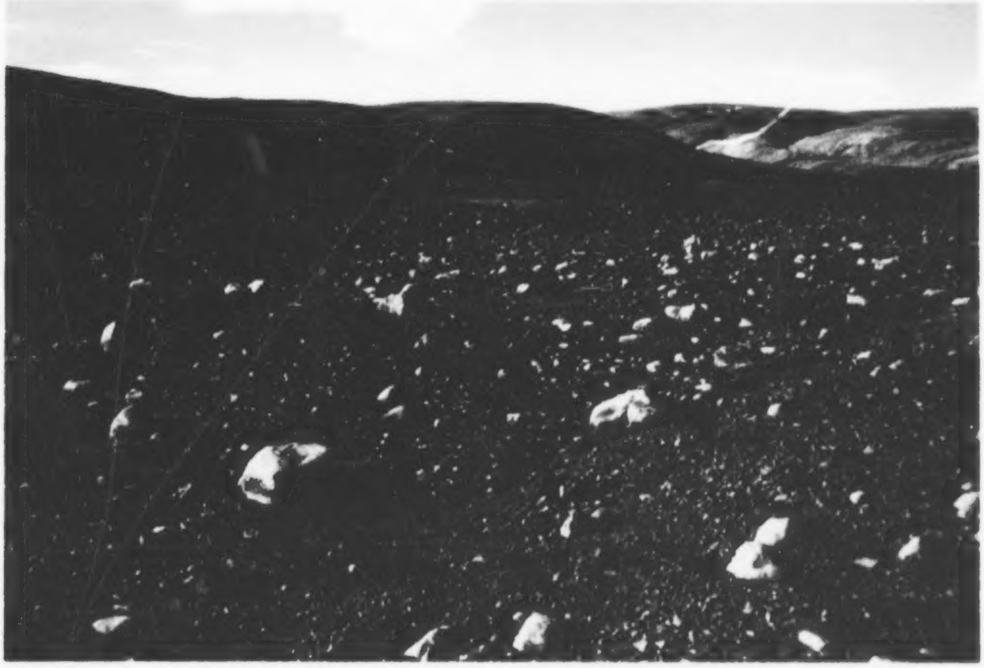


Figure 3.13: Till veneer on Weather station lowland. The dashed line represents a lithological boundary in the steeply dipping bedrock strata which is mimicked on the till surface.

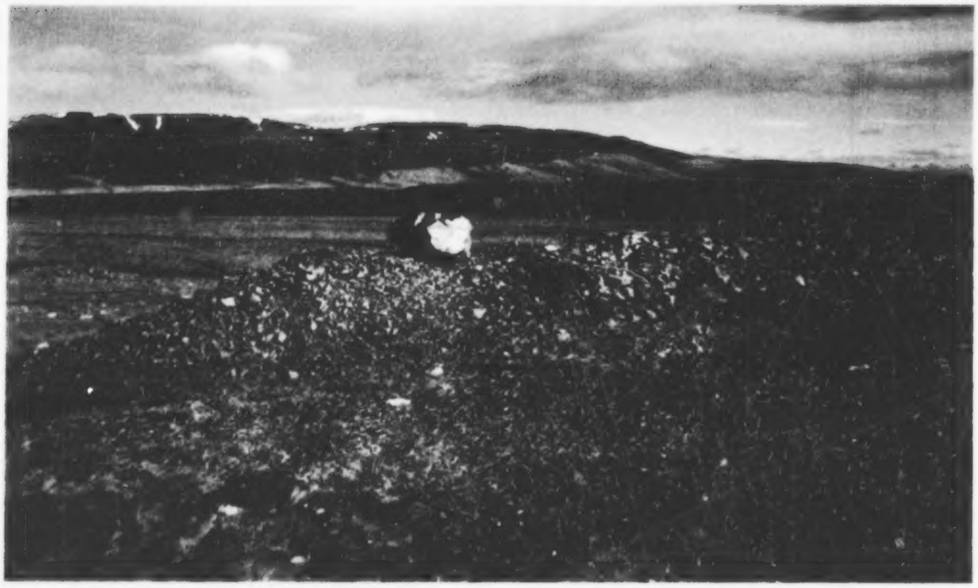


Figure 3.14: Discontinuous till veneer on the eastern slope of Troelsen ridge. The glacial erratic rests on a sandstone ridge with weathered shale in foreground. View towards Black Top Ridge across Weather station lowland.

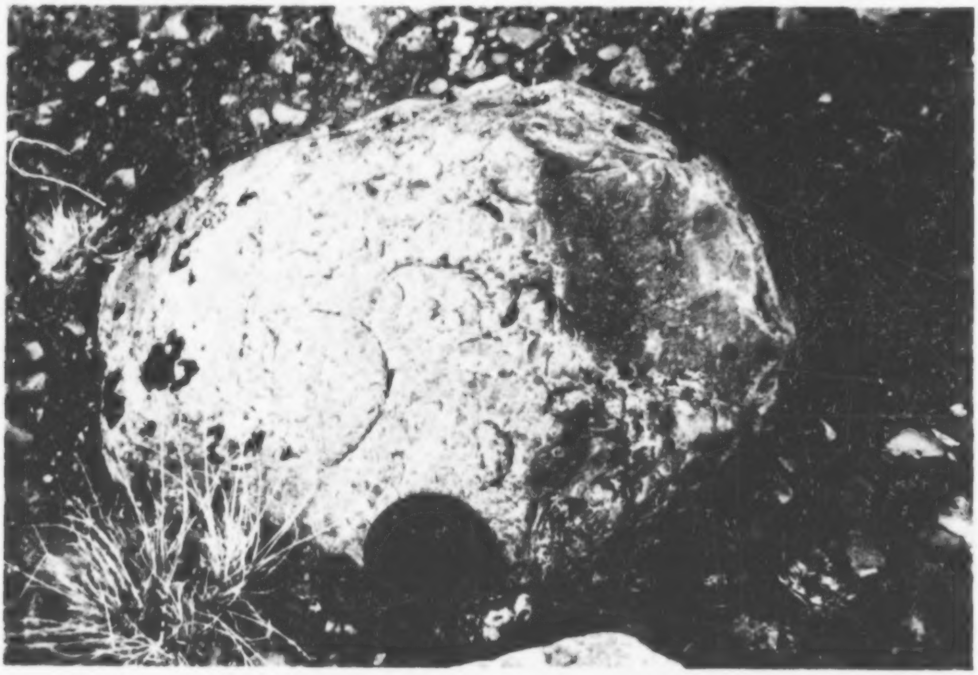


Figure 3.15: Typical appearance of quartzite erratics in the study area, displaying crescentic scars like percussion or chatter marks.

Peninsula, then quartzite erratics would be polycyclic and inappropriate as ice-flow indicators.

It may be argued that granite erratics are also primarily fluvial in origin because they share a common source area with quartzite in the Canadian Shield of southeastern Ellesmere Island. However, granite boulders were up to 3 m in diameter, too large to be fluvial in origin and their restricted distribution along the northwestern fringe of the study area is more typical of glacial dispersal (Figs. 3.16 and 3.17). Consequently, the distribution of granite erratics must reflect ice trajectories and their presence or absence forms the basis for describing till distribution in the study area.

3.3.2 Distribution

(A) Till with granite:

Iceberg Point lowlands

A fine-grained till containing granite erratics blankets the coastal lowlands around Mt. Lockwood between Iceberg Point, Snow Goose hill and Eastwind Lake (Figs. 2.14 and 3.1). The upper limit is indistinct; till is generally absent above 275 m asl on the northern slope of Troelsen ridge, it is present at 200 m asl northwest of Eastwind Lake, and it drapes the western slope of Snow Goose hill up to 160 m asl. The concentration of granite erratics decreases from west to east across the lowlands such that only one clast was observed between Eastwind Lake and Greely Fiord and its occurrence below Holocene marine limit may imply an ice-rafted origin. (Granite boulders were observed on the modern shore of Greely Fiord.) Another characteristic of this till is its fossiliferous nature. Whole and fragmented shells were observed both in section and as a lag on till surfaces (sites[1] 26-30).

In section, the till was composed of a hard, fine-grained matrix with striated erratics (including sandstone, conglomerate, quartzite and granite; sections 2, Appendix B). A slump along a river south of Mount Lockwood exposes at least 3 m of till (Fig. 3.18) whereas Hodgson and Edlund (1977) reported till thicknesses of 15 m in this area.

Hare Cape ridge

Abundant granite erratics and fragmented shells are scattered over weathered bedrock on Hare Cape ridge. In places, shelly till veneer containing granite is preserved on low gradient slopes and topographic lows. This fossiliferous diamicton was

---

[1] The site number refers to a location on Figure 3.10 where shells were collected for amino acid analysis. The results of this analysis can be found on Table C.2 also using this number; however, their significance is discussed in the following chapter.

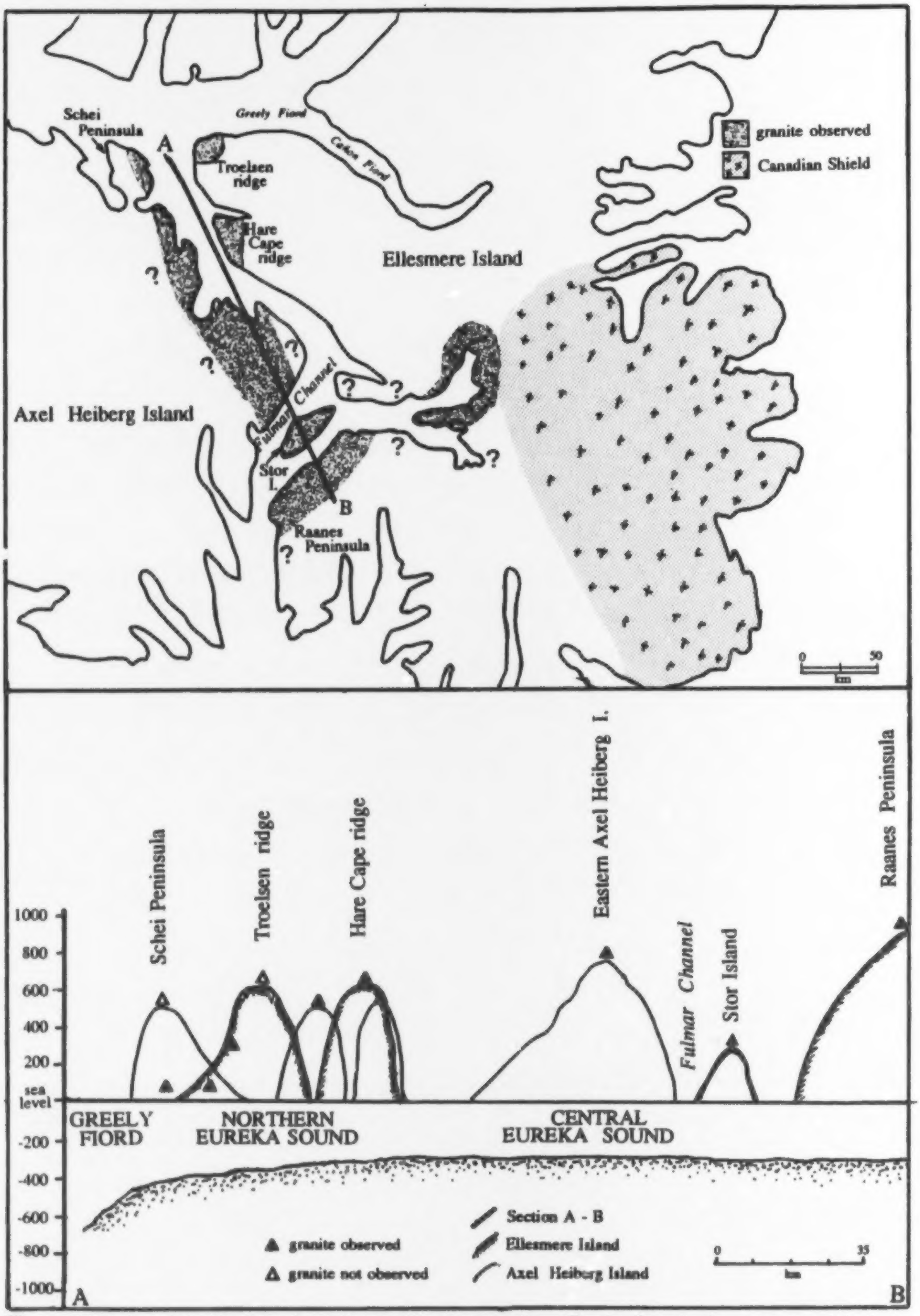

Schei Peninsula
Greely Fiord
Cañon Fiord
A
Troelsen ridge
Hare Cape ridge
Ellesmere Island
granite observed
Canadian Shield
?
Axel Heiberg Island
Fulmar Channel
Stor I.
B
Raanes Peninsula
0 50 km
Schei Peninsula
Troelsen ridge
Hare Cape ridge
Eastern Axel Heiberg I.
Fulmar Channel
Stor Island
Raanes Peninsula
1000
800
600
400
200
sea level
-200
-400
-600
-800
-1000
GREELY FIORD
NORTHERN EUREKA SOUND
CENTRAL EUREKA SOUND
granite observed
granite not observed
Section A - B
Ellesmere Island
Axel Heiberg Island
0 35 km
A
B

Figure 3.16: *Top:* Distribution of granite erratics along northern Eureka Sound. ? = area not visited. *Bottom* Profile AB through Eureka Sound showing upper limit of granite erratics.

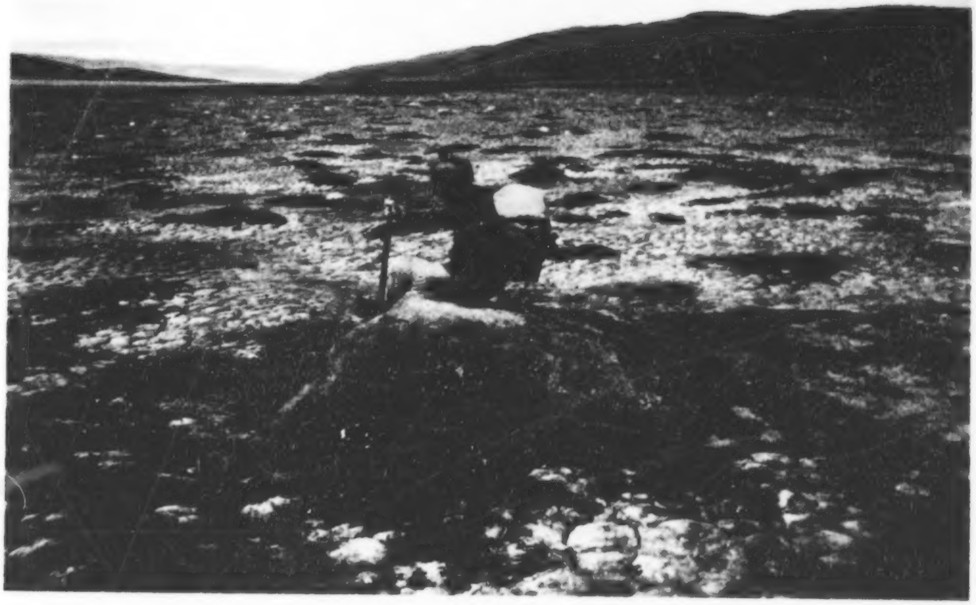


Figure 3.17: Large granite boulder near Blue Man Cape. Granite clasts are common here forming part of a dispersal train along northern Eureka Sound.

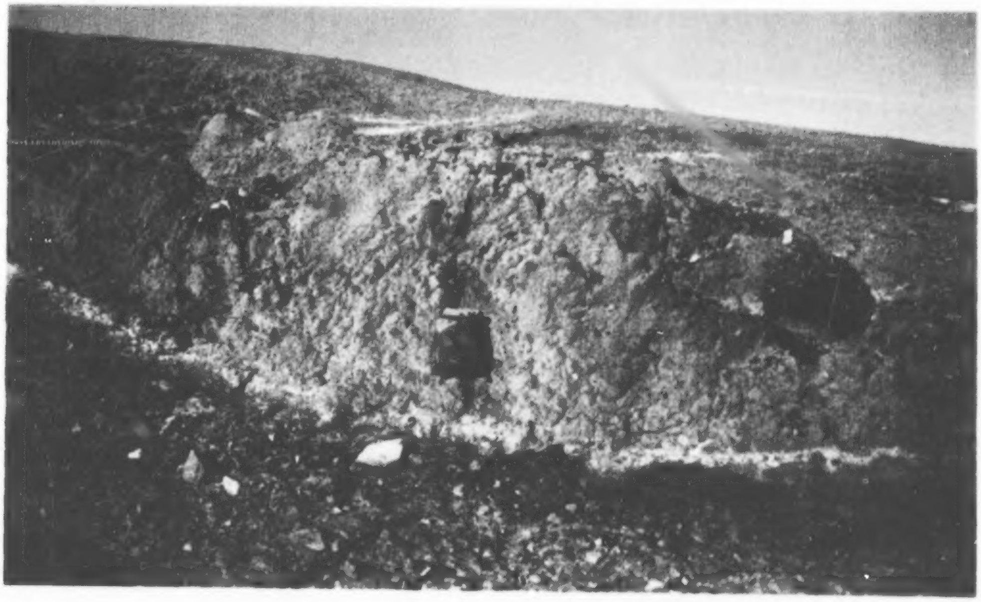


Figure 3.18; Slumped exposure of granite-rich till on Iceberg Point lowland. The till has a hard, fine-grained matrix containing striated erratics and shell fragments.

previously described by Sim (1961). He favoured glacial transport of shells to the ridge summit although he also entertained the possibility that Hare Cape ridge was once submerged below sea level. Conventional radiocarbon dates on some shell samples provided ages of 30 300 ± 1600 BP (outer fraction) and 36 300 ± 2000 BP (inner fraction, GSC-111, 625 m asl; Dyck and Fyles 1964), 28 700 ± 600 BP (GSC-51, 192 m asl; Dyck and Fyles 1962), and 19 500 ± 1100 BP (L-548, 607 m asl; Sim 1961). The age differences between inner and outer fractions of sample GSC-111 prompted Fyles (*in* Dyck and Fyles 1964, p. 178) to suggest that these dates are only minimum estimates and that all the samples were likely glacially transported. Shell samples from selected elevations were collected for radiocarbon analysis (sites 33-36).

Romulus lowland

Till blanket, veneer and discontinuous veneer were recorded on Romulus lowland. Granite erratic concentrations were low (≤1%) relative to those recorded on adjacent Hare Cape ridge and were only observed in the western one-third of the lowland. The shell content of the till decreased with distance from Eureka Sound and valves became more fragmented inland (site 38).

A section along a tributary valley of Romulus river* on the northern lowlands revealed interbedded tills and fan/deltaic sediments (section 4, Appendix B). In general, the stratigraphy indicates two discrete glacial episodes. The two lower tills are interbedded with sandy units that show a distal to proximal transition in a prograding subaqueous environment. The fan/delta sequence indicates substantial crustal depression (90-100 m) during glacial buildup prior to arrival of ice from Eureka Sound. The uppermost till occurs discontinuously on the lowlands and represents the most recent glaciation of the area.

Central Eureka Sound

On southwestern Fosheim Peninsula granite erratics were observed only below marine limit along the Eureka Sound coast. Occasional granites were found on a washed diamicton surface which also included striated sandstone clasts and abundant whole and fragmented shells. Other shell samples from similar elevations provided Holocene dates. Although granites may have been ice- rafted during higher Holocene sea levels, there remains the possibility that a till containing granite underlies the marine veneer in this area. Farther north, till outcrops above marine limit; however, no granites were observed although fragmented shells were present (see next section).

On Stor Island, angular granite boulders were observed near the summit at 535 m asl which corroborates earlier reports by Troelsen (1952). Granite blocks and associated fine-grained, shelly till were mapped on high ground (470 m) above Fulmar Channel on eastern Axel Heiberg Island (site 37; Fig. 3.16). Farther north, granite erratics rest on weathered summits above Depot Point (725 m) and Fair Cape (455 m). Rudberg (1963) reported granite erratics on Schei Peninsula but their elevation is uncertain. Meanwhile, granites were not observed during foot traverses to local summits (450 m asl) on the northern tip of the peninsula in 1990. Similarly, across Flat Sound on northern Axel Heiberg Island, J. Bednarski (pers. comm. 1989) recorded only local lithologies in surficial deposits.

The distribution of granite erratics along Eureka Sound is only roughly known considering the low density of sampling points and the confidence associated with the absence of evidence at particular sites. Nevertheless, the regional pattern shows a long and relatively narrow belt of glacial debris with a distinctive granitic composition. This distribution is typical of a *Dubawnt*-type dispersal train where debris is spread down-ice from a relatively restricted source area (Dyke and Morris 1988). The nearest source of granite is 120 km southeast in the northernmost Canadian Shield on Ellesmere Island; however, other sources may include northwestern Greenland or the Canadian Shield of the central Arctic.

(B) Till without granite

West-central Fosheim Peninsula

For the most part, till on Troelsen ridge, Black Top Ridge and the adjacent lowlands consists of a discontinuous veneer dominated by quartzite and sandstone erratics (Fig. 3.1; see Figs. 3.13 and 3.14 for examples). Local depressions on Troelsen and Black Top ridges contain weathered till blanket with large quartzite, sandstone and diabase clasts in a fine-grained matrix (outcrops are too small to be shown on Fig. 3.1). Erratic blocks of sandstone were observed below the summit of northeastern Black Top Ridge.

Central Eureka Sound

Weathered bedrock and felsenmeer, including occasional quartzite and striated sandstone erratics, were documented on the summit of Airfield ridge (672 m); however, neither granites nor shell fragments were observed. Farther south, a fossiliferous, stony till mantled the lower slopes of the Sawtooth Range overlooking Eureka Sound (sites 32 and 40).

Central lowlands

Till blanket and veneer cover extensive parts of the central lowlands of western Fosheim Peninsula (Fig. 3.1). For the most part, this till consists of a silty matrix with abundant quartzite, carbonate, sandstone and diabase erratics. Where the till rests on Neogene terrace sediments, it is characteristically coarser with a more gravelly appearance. Although winnowing of fines from the till surface contributes to its coarser appearance, the till in section retains its high gravel content supported by a fine-grained matrix (section 1, Appendix B). The increased gravel content of till overlying sandy terrace sediments is difficult to explain unless, as suggested previously, some gravel has been incorporated locally into the till.

Fine-grained till mantles extensive areas adjacent to central Cañon Fiord and along the foothills of the Sawtooth Range whereas only isolated patches are preserved on interfluves and valley bottoms east of Black Top Ridge (Fig. 3.1). Fossiliferous till was observed at only one site on the lowlands, about 5 km southwest of Cape Lockwood (site 39). The shelly, fine-grained till blanketed a small area and could not be traced in any direction. There is a possibility that local gravel terraces overlie the till but there is no stratigraphical evidence to substantiate this.

Sections in the fine-grained till are restricted to Slidre River valley where local incision has provided some stratigraphical record of glacial, glaciomarine and fluvial events. Stony till overlying bedrock is exposed discontinuously along the eastern valley rim. The till probably forms a continuous unit because slumped debris mantling the valleyside is composed of a similar diamicton, including large striated erratics. The upper surface of the till is commonly mantled with marine silts containing *in situ* shells of Holocene age.

Near Hot Weather Creek a retrogressive thaw slump exposed frozen till in its 3 m high headwall (section 5, Appendix B). The till displays evidence of subglacial tectonic shear, including stacking and rotation of sub-till sand rafts and attenuation of sand laminae in the direction of shear. The orientation of shearing suggests former ice movement from the east or southeast. The till is draped by a thin sand unit and Holocene marine silt but no conclusion can be made regarding the age of the till.

3.3.3 Associated landforms

Iceberg Point

The coastal lowlands east of Iceberg Point are blanketed with till which is mostly featureless. However, near Iceberg Point the till plain has an undulating surface with rounded ridges and shallow depressions (Figs. 3.19 and A.5). Although these features are easily recognizable on air photos, they are less impressive on the ground. Ridge crests (*ca.* 130 m asl) have little or no gradient and are mantled with fossiliferous till similar to the surrounding region. Their light tone on air photos reflects the absence of vegetation. Depressions between ridges and hummocks contain accumulated organics, low-centered polygons and shallow ponds. The eastern edge of this topography is marked by a subdued, sinuous ridge which has a general north-south orientation. The area lies below the limit of Holocene submergence and therefore, surficial materials have been reworked by marine processes.

The ridges are interpreted to be moraines or former grounding lines constructed during deglaciation of the lowlands. Till lithology indicates a former advance of ice onto the lowlands from Eureka Sound. The limit of this advance is marked by the distribution of granite erratics which are present up to 275 m asl above Iceberg Point but were not observed above marine limit farther east.

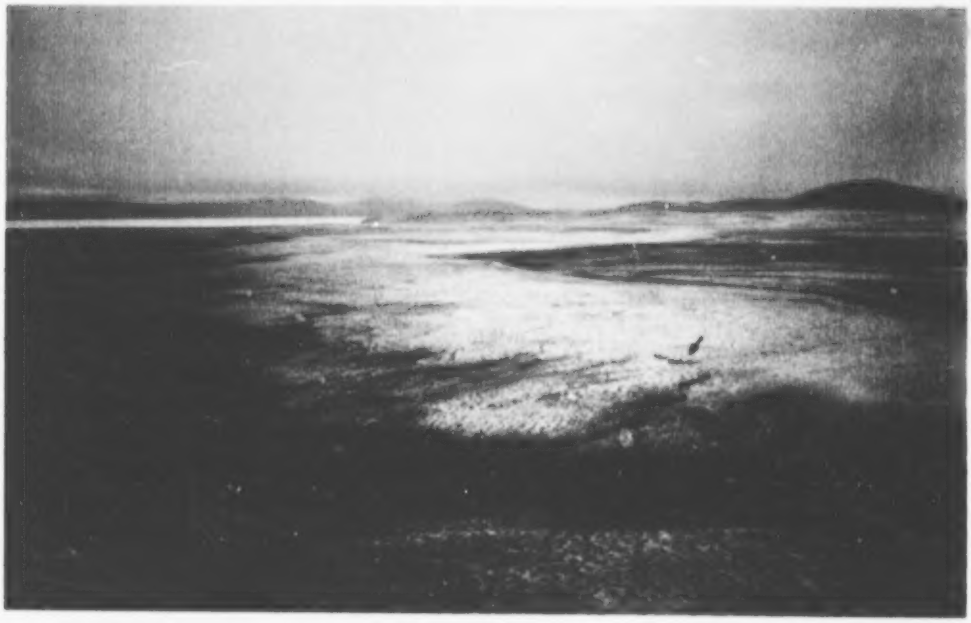


Figure 3.19: Oblique view of marine-washed moraine or grounding line near Iceberg Point. Mount Lockwood (right) and Greely Fiord (left) in background. A herd of muskoxen provides scale.

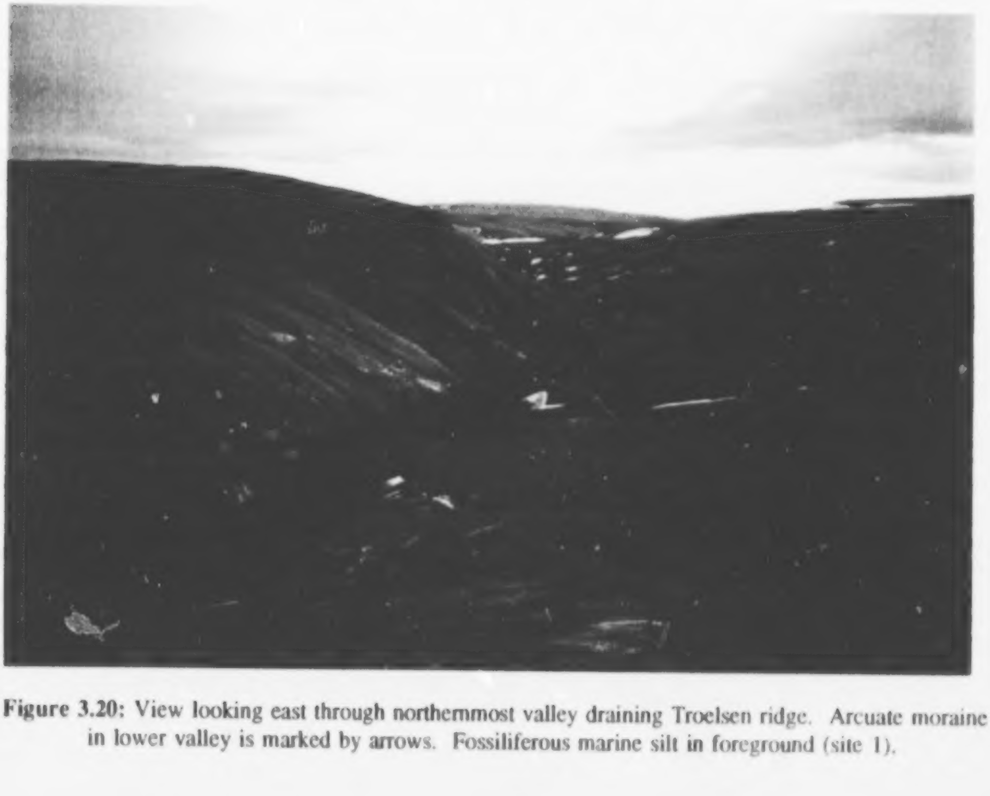


Figure 3.20: View looking east through northernmost valley draining Troelsen ridge. Arcuate moraine in lower valley is marked by arrows. Fossiliferous marine silt in foreground (site 1).

Troelsen ridge

Four main rivers drain the western side of Troelsen ridge and flow into Eureka Sound. The northernmost river valley cuts through the ridge as a deep gorge and consequently, has the largest drainage basin. A small arcuate moraine occupies the lower valley (*ca.* 100 m asl) where it faces onto Eureka Sound (Fig. 3.20). Beyond the moraine dissected outwash overlies fossiliferous marine silt. The moraine was deposited by a valley glacier descending from adjacent Troelsen ridge.

Eastwind Lake lowlands

The lowland between Eastwind Lake and Snow Goose hill is mantled with fossiliferous till; however, adjacent to Black Top Ridge alluvial gravel and coarse, non-fossiliferous till predominate. The till consists of local lithologies from Black Top Ridge and in places, is bordered by moraines (Fig. 3.21). Both were deposited by ice which advanced onto the lowland through short, steep-sided valleys on western Black Top Ridge. The location of moraines and the distribution of local till suggest that ice extended at most 2 km onto the lowland, probably as thin piedmont lobes. Moraines descend from 230 m to 170 m asl onto a till and alluvial plain. There is no evidence that these former outlet glaciers contacted a higher sea level.

Hare Cape ridge

Overfit channels outline the drainage of a former ice cap on Hare Cape ridge. A former ice margin is marked by a lateral moraine on the eastern side of a small valley that drains the central part of the ridge (Fig. 3.22). The lower limit of the moraine is 120-130 m asl and this may represent the maximum ice position where it contacted the sea. Downvalley, fossiliferous, stony marine silt at 75 m asl indicates ice-proximal sedimentation into a higher relative sea level.

Romulus lowland

Romulus lowland contains one of the most significant glacial landforms in the study area. It is dominated by a 25-30 km long sinuous ridge which runs diagonally across the lowland to the Slidre River (Figs. 3.1 and A.6). For the most part, the ridge consists of a single, well-defined crest but in places appears multiple-crested (Fig. 3.23). Ridge relief above adjacent bedrock varies between 12 m and 20 m. The ridge is interpreted to be a moraine (Romulus moraine) deposited by ice advancing north across the lowland. The southern, ice-proximal side is clearly defined by a sharp bedrock-moraine contact whereas the northern margin is more difficult to identify because it is draped by stony, marine silt.

The moraine crest declines in elevation from 134 m asl in the southwest to 114 m asl near Romulus Lake. Farther east the moraine is discontinuous and appears to bifurcate; one ridge continues east rising from 78 m to 87 m and becomes indistinct near Big Slide creek; the other ridge veers northeast descending to 57 m along Slidre River where it is buried by Holocene marine sediments (Fig. A.7).

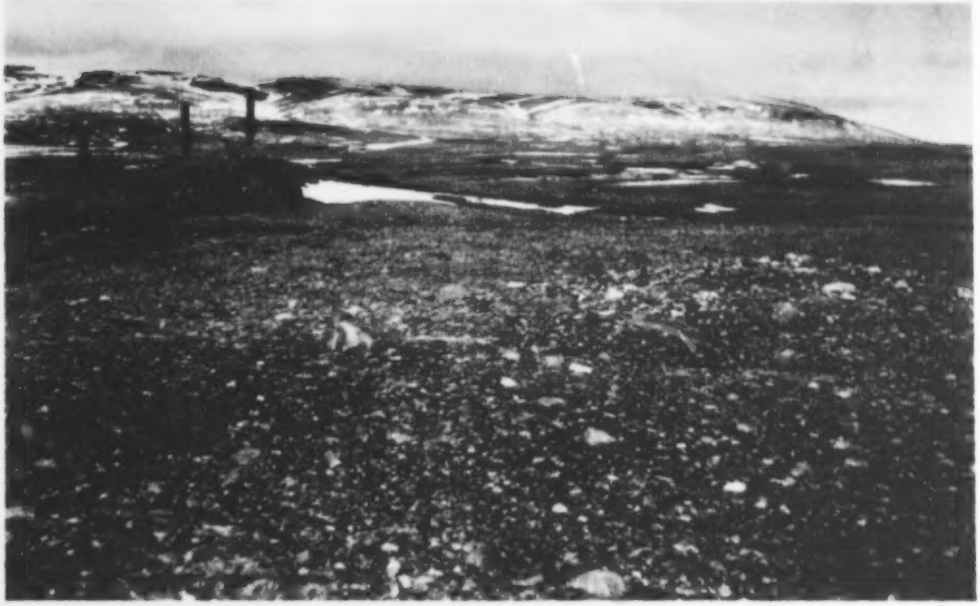


Figure 3.21: View looking west across till and an end moraine (arrowed) deposited by local piedmont glaciers from Black Top Ridge. Troelsen ridge is in the background.

Figure 3.22: View looking upvalley at moraines (arrowed) deposited by an outlet glacier from a former ice cap on Hare Cape ridge.

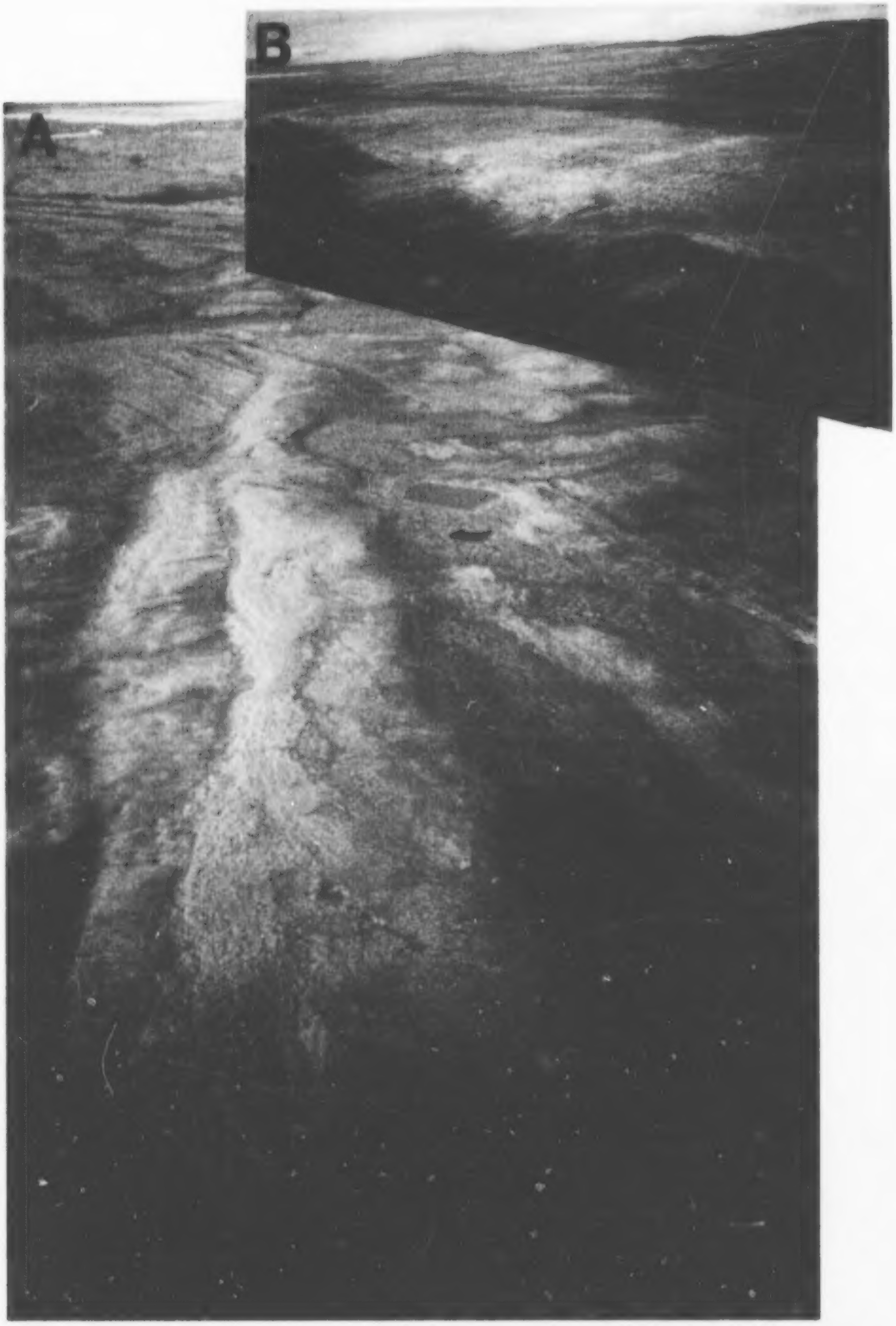


Figure 3.23: A. Oblique view of Romulus moraine looking southwest towards Eureka Sound. The proximal side is to the right. Bedrock is exposed on the lowland near the end of the moraine. Beach berms demonstrate marine washing of the moraine surface. **B.** Oblique view of Romulus moraine from its proximal side.

Surficial material on ridge crests consists of a gravel lag with frequent subrounded boulders. In places, these boulders have been used in the construction of stone rings and caches (Fig. 2.3). Because the moraine lies below the limit of Holocene marine submergence, there has been some winnowing of fine-grained sediment and reworking of gravel into beach berms (Fig. 3.23). Pockets of Holocene marine silt drape some ridges to within 10-15 m of their crest. At its southwestern extremity, the gravel lag is discontinuous exposing fossiliferous, fine-grained diamicton. The diamicton is an extension of the till that mantles the adjacent lowland.

Along Slidre River, there is an abrupt change in sediment where the buried moraine is exposed along the valley sides. Holocene marine rhythmites bracket a 17 m section of diamicton consisting of a partially indurated, structureless, fine-grained matrix with abundant striated clasts, including carbonate, sandstone and quartzite. The river channel below the section is armoured with large, striated boulders which form a knickpoint in the river profile (Fig. 3.24). These boulders have been eroded from morainal sediments but are too large to be transported by Slidre River.

There is no surface expression of Romulus moraine north of Slidre River; however, till exposed along the eastern valley rim may represent its extension. Ice-directional indicators in the till suggest a westward or northwestward advance of ice. Moreover, with distance northeast along Romulus moraine, surficial sediments become non-fossiliferous and the proportion of carbonate clasts increase resembling the till on the central lowlands. Therefore, it is concluded that Romulus moraine was deposited both by ice advancing north across Romulus lowlands from Eureka Sound and by ice advancing west or northwest across the central lowlands of Fosheim Peninsula.

Central Eureka Sound

A sharp-crested moraine occurs at 225 m asl on the southern slope of the Sawtooth Range above Eureka Sound (Fig. 3.25). The ridge is *ca.* 150 m long with a local relief of 20-30 m. Surficial deposits consist of a bouldery lag with abundant sandstone and some quartzite clasts. Because the moraine is oriented parallel to Eureka Sound, it was likely deposited by ice occupying the sound. Fossiliferous, stony till is preserved on adjacent slopes above 230 m asl and consequently, the moraine is likely recessional.

All of the valleys visited along this part of the coast contain evidence of former ice draining from the Sawtooth Range towards Eureka Sound (Fig. 3.26). Short arcuate moraines descend from 220 m to 80 m asl and in some cases, local ice has overridden fossiliferous till deposited by the regional ice from Eureka Sound (Fig. 3.26).

Central Lowlands

Meltwater channels eroded in till on the central lowlands are characteristically wide and shallow with flat floors covered by a gravel lag (Figs. 3.1 and 3.27). They are easily recognizable on air photos because they contain strings of shallow ponds

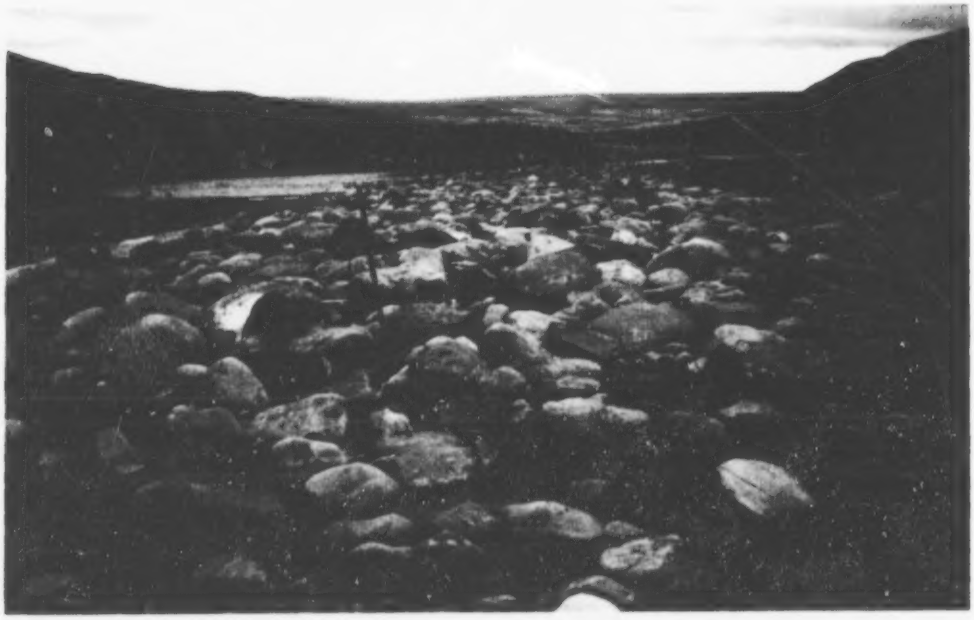


Figure 3.24: Boulders eroded from Romulus moraine armour Slidre River channel.

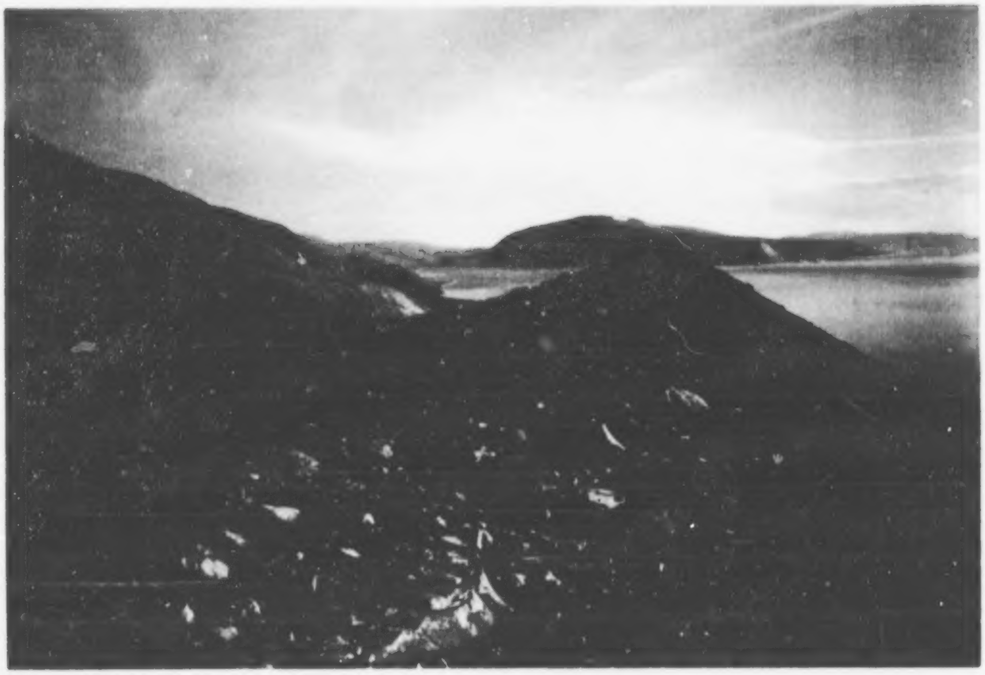


Figure 3.25: Photograph of lateral moraine at 225 m asl oriented parallel to Eureka Sound (in background).

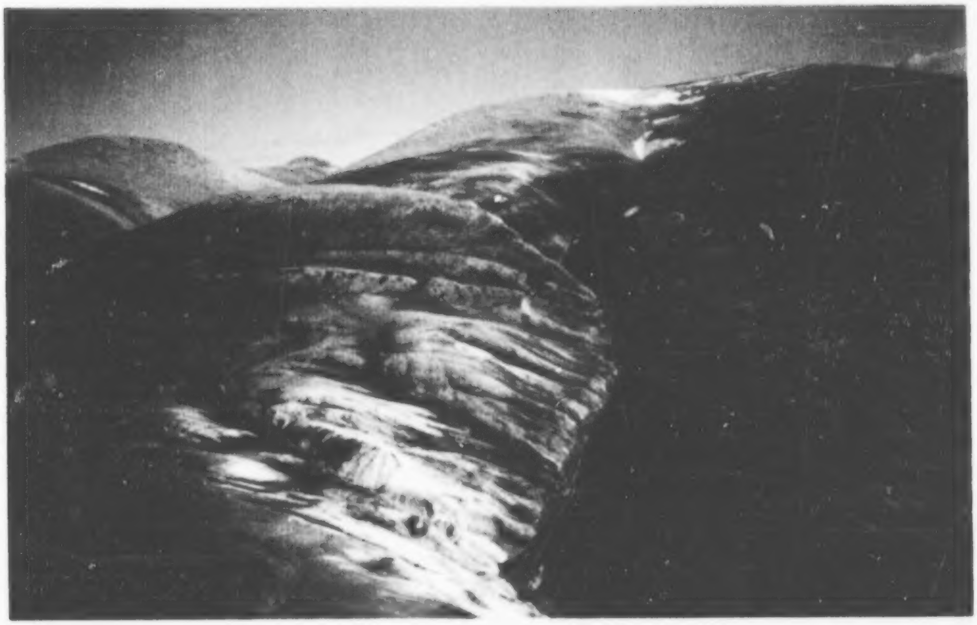


Figure 3.26: View looking west towards the Sawtooth Range where it intersects Eureka Sound. Moraines (arrowed) mark the former margins of an outlet glacier descending through the valley from the ridge summit (background). **RT** denotes regional till overridden by local ice.

Figure 3.27: Coarse gravel lag on the floor of meltwater channels eroded in till on the central lowlands.

reflecting low gradients and show a marked increase in vegetation due to moisture availability.

The till plain between Slidre River and Cañon Fiord displays a radial pattern of meltwater channels with some exiting into former marine embayments of Cañon Fiord and others terminating in Slidre River valley (Figs. 3.1 and A.3). Raised marine sediments are present at channel mouths and usually mark the limit of Holocene marine submergence. The origin of the meltwater associated with these channels is problematic. The channels are located 25 km west of the Sawtooth Range and therefore, it is unlikely that local cirque glaciers would have advanced such a distance. Instead, it is suggested that a local ice cap developed along the foothills of the Sawtooth Range in conjunction with the growth of cirque glaciers. Meltwater from the western margin of this ice cap would have fed channels on the central lowlands.

3.4 GLACIOFLUVIAL SEDIMENTS

Glaciofluvial sediments constitute a relatively small proportion of the surficial deposits on western Fosheim Peninsula. Although in some cases, the classification of these deposits was straightforward, in others it was difficult to distinguish between gravelly outwash deposits and Neogene terrace sediments overlain by gravelly diamicton. As a result, only those areas clearly covered by glaciofluvial material were classified as such on the surficial geology map (Fig. 3.1).

3.4.1 Distribution

Weather Station lowland

The lowland between Eureka weather station and Eastwind Lake is largely composed of weathered bedrock with a discontinuous till and marine sediment veneer. In marked contrast, however, a locally prominent ridge runs obliquely across the lowland from southwest to northeast (Fig. 3.28). The ridge occupies the crest of an elongate bedrock high that forms the central drainage divide between Station and Black Top creeks. Where the ridge is continuous and well-defined, it has an elevation between 176 m and 185 m asl. Shorter ridge segments and flat terraces occur at lower elevations (165-170 m asl) farther north and show evidence of marine washing. Therefore, the local marine limit is considered to be 170 m asl.

Sections were not observed in this landform but its surficial material is similar to the gravelly diamicton that overlies Neogene terraces. Considering its topographic position, this landform may also be an erosional remnant of high terrace sediment, protectively capping less resistant bedrock. However, the narrow, sinuous form of the ridge is unlike the terrace morphology of Neogene sediment remnants observed

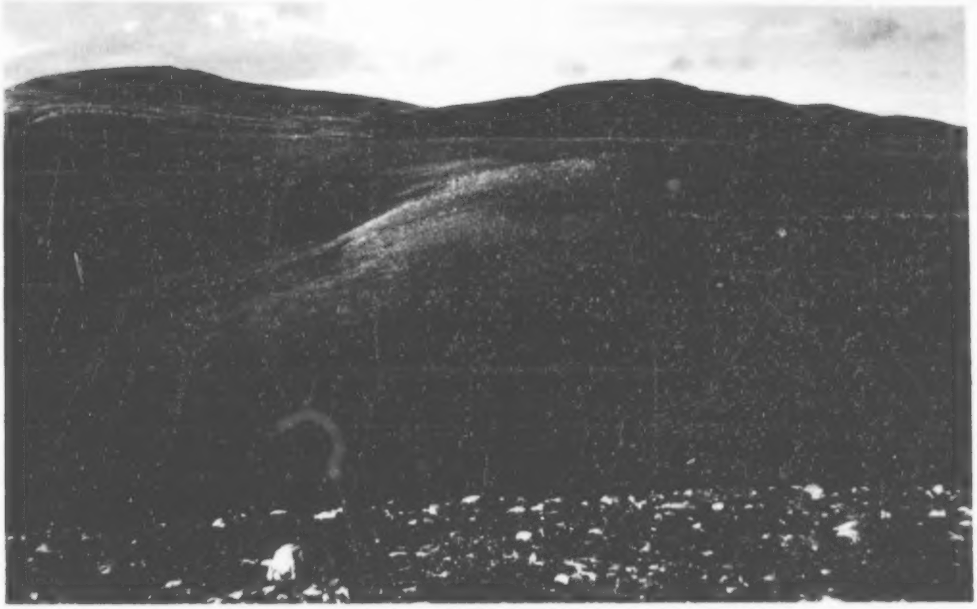


Figure 3.28: View looking north across discontinuous kame moraine (185 m asl) on Weather station lowland. Contrast with Neogene terrace remnants in Figure 3.7.

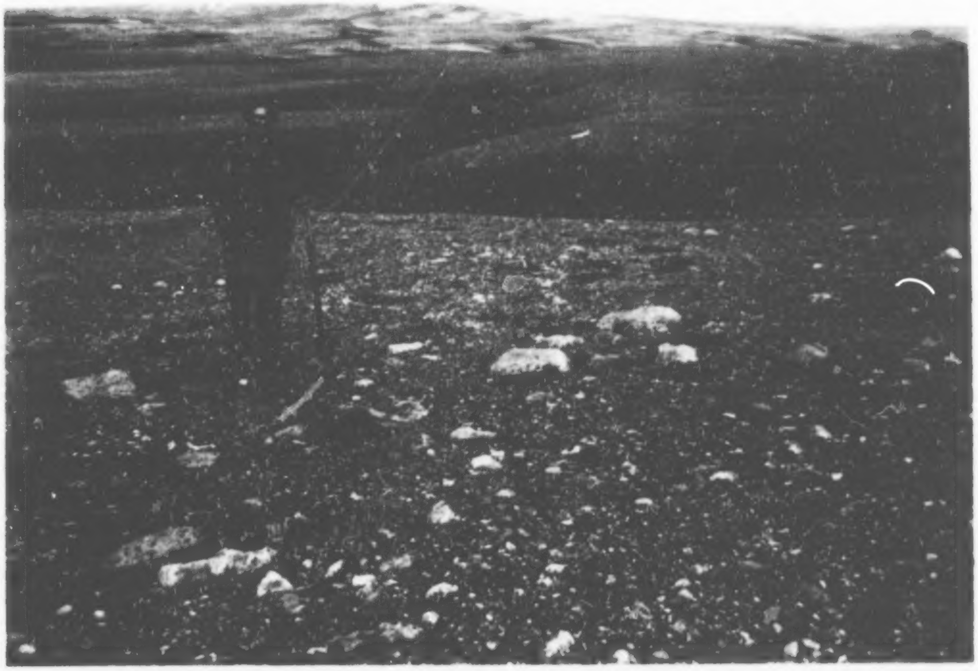


Figure 3.29: Dissected surface of coarse delta deposits at *ca.* 170 m asl at the northern end of Black Top Ridge. Similar gravel terraces at lower elevations grade north towards Greely Fiord.

elsewhere on the peninsula. Consequently, I propose that the ridge has a glaciofluvial origin, deposited as a kame moraine during deglaciation of the lowland.

Black Top Ridge and Snow Goose hill

The lowland between Black Top Ridge and Snow Goose hill is dominated by gravel outwash overlying fossiliferous till. Discontinuous gravel ridges (170 ± 3 m asl) are interpreted to be the remnants of a kame moraine from which outwash terraces grade (Figs. A.1, A.2 and 3.29). Occasionally, large sandstone blocks were observed on ridge crests and, where former meltwater has dissected the ridge, numerous well-rounded quartzite and sandstone boulders armour the channel bed. Outwash terraces grade west towards Eastwind Lake lowlands and northeast towards Greely Fiord; however, there is little continuity to these surfaces which have likely been reworked and regraded to lower base levels as ice retreated from the basin. The outwash terraces are considered to be the remnants of a dissected kame delta that was deposited into a higher sea at *ca.* 170 m asl while ice stood at the moraine.

At the northern end of Black Top Ridge, small outwash fans occur where steep meltwater channels intersect the lowlands. Outwash originating from Black Top Ridge is characteristically darker on air photos than that farther east because it contains a higher content of local shale and diabase. These fans grade to various elevations but the lowest surfaces clearly terminate at 148-149 m asl; hence, Holocene marine limit.

Meltwater channels on Snow Goose hill contain outwash material where they descend onto the surrounding lowlands. On the northern side of the hill, meltwater from the largest channel deposited an outwash fan at 152 m asl. Downslope, outwash has been reworked into beach berms marking marine limit at 148 m asl. Farther southwest, outwash occupies the lower outlet of meltwater channels descending from Snow Goose hill onto Eastwind Lake lowlands.

Big Slide Creek

Near the head of Big Slide Creek, a narrow, outwash plain grades downvalley from 232 m to 185 m asl (Figs. 3.1 and 3.30). The sediments consist of pebble to boulder gravel more than 5 m thick, resting on poorly consolidated sandstone and coal of the Eureka Sound Group. Paleochannels are preserved on the outwash surface at least 20 m above the modern creek bed (Fig. 3.31). The outwash plain was deposited by proglacial meltwater streams originating from an ice source on the plateau at the head of Big Slide Creek. The lower outwash surfaces which deflect the modern stream may represent a degraded delta. Gravel ridges on a flat bench at 184 m asl are tentatively interpreted to be beach berms and sea ice-pushed ridges marking a former sea level at 184 m asl.

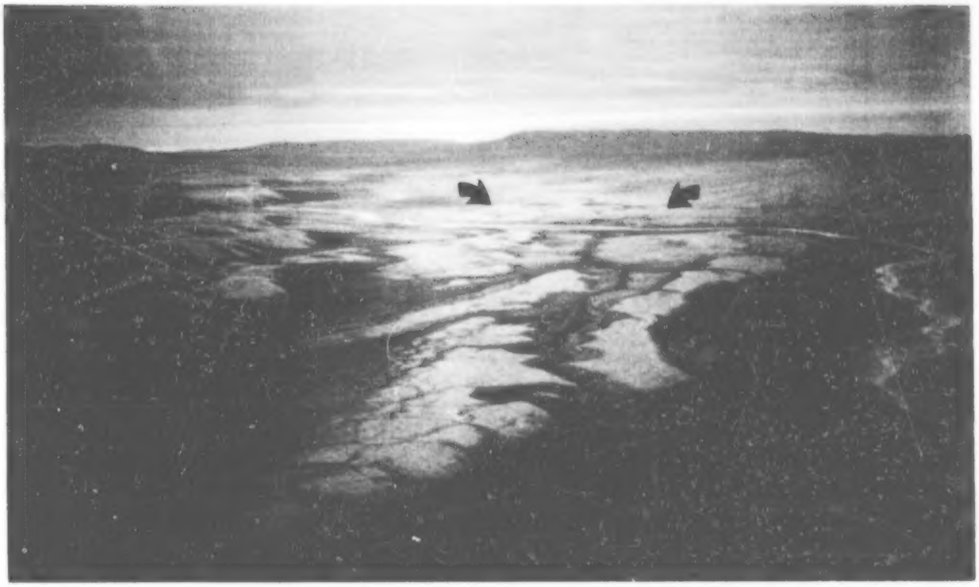


Figure 3.30: Dissected outwash above Big Slide creek grades downvalley to a degraded delta at *ca.* 185 m asl (arrowed). The delta deflects the modern creek channel.

Figure 3.31: Paleochannels preserved on the outwash surface in Big Slide creek valley. They are *ca.* 10-15 m wide and 3-5 m deep.

3.5 RAISED MARINE SEDIMENTS

Raised marine sediments have been divided into two subunits: (1) deltaic, beach, and nearshore deposits, and (2) marine sediment veneer and washed bedrock. Together they form the third largest surficial unit in the study area (Fig. 3.1). Glaciomarine deposits are identified on the basis of their former relationship to a receding ice margin and, in general, consist of thicker accumulations relative to postglacial deposition.

3.5.1 Deltaic, beach, and nearshore sediments

Extensive *raised marine deltas* are rare in the study area which may reflect either a deficiency in sediment entering the sea (England 1983) or the poor preservation of these landforms in modern river valleys. However, preservation has not been a noted problem in other high arctic areas due to the regional aridity which greatly limits erosion rates. The most common raised delta type on the peninsula consists of thin accumulations of sandy sediment where meltwater entered former marine embayments. Meltwater channels are commonly dry or contain underfit streams which are presently unable to remove the deltaic sediment. Coarse delta remnants are preserved in formerly glaciated basins of the Sawtooth Range where steep slopes and seasonal meltwater production favoured episodically high discharges capable of transporting cobble and boulder-size clasts (Fig. 3.32).

Raised beaches are generally composed of sand and gravel reworked from underlying deposits. They are best preserved at lower elevations which likely reflects the lower rates of emergence during the last 4-5 ka. In some areas, however, prominent beaches were observed almost to the limit of marine submergence, demonstrating that local coastal processes were independent of emergence rates. Sea-ice pushed ridges are common on modern beaches, especially where there is exposure to wide marine channels, allowing wind-driven ice movement onshore (Fig. 3.33). Similar ridges occur above present sea level.

The abundance of *nearshore sediment* on the peninsula supports the argument for predominantly suspended sediment in the deglacial meltwater load. Where this sediment was observed in section, two main lithofacies were identified: lithofacies **Ia** consists of horizontally and thinly bedded mud with interlaminated to interbedded sand (Fig. 3.34). These rhythmites are interpreted as suspension deposits or cyclopels where particle deposition in discrete laminae results from differential settling during the progressive decay of a turbid overflow plume (Domack 1984; Mackiewicz *et al.* 1984; Cowan and Powell 1990; Stewart 1990).
Lithofacies **Ib** consists of multiple graded beds of fine sand to silt (Fig. 3.35). They are interpreted to be cyclopsams derived by suspension settling from turbid plumes. Together with the cyclopels, these sediments represent a distal to proximal change in

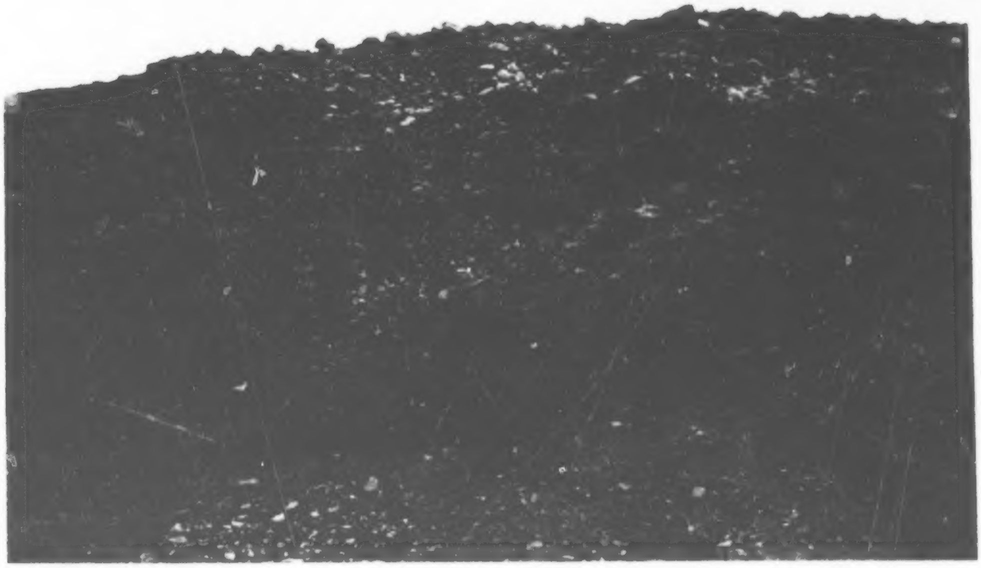


Figure 3.32: Delta foresets inclined at 17-22° are composed of clast-supported gravel and coarse sand (site 18). Horizontally-bedded gravel (topsets?) rests unconformably on the foresets.

Figure 3.33: Sea-ice pushed ridge composed of sea ice and fine-grained sediment at Iceberg Point. Smaller ice-free ridges are indicated by arrows.

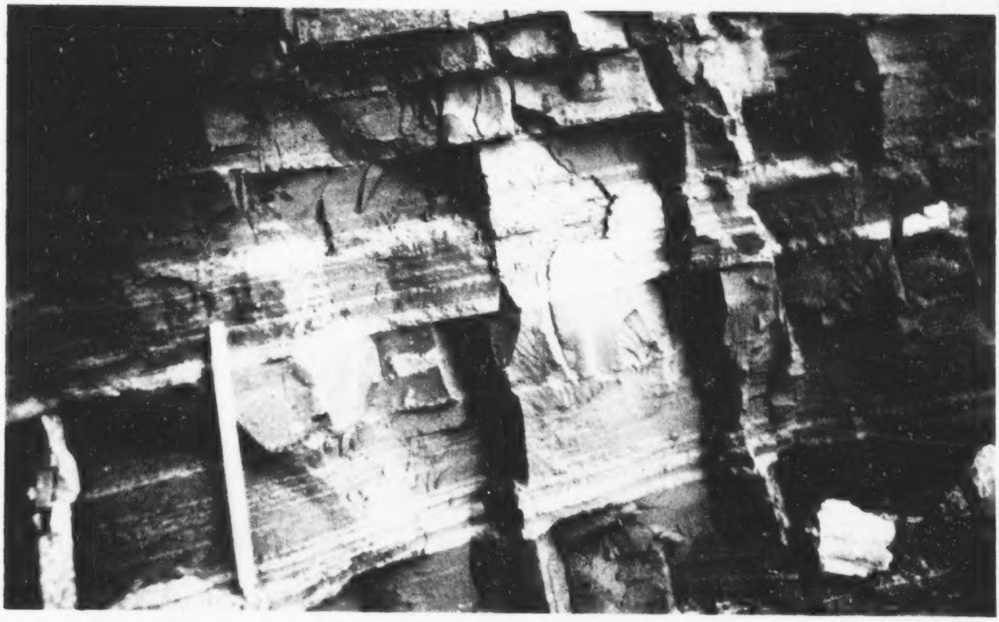


Figure 3.34: Lithofacies **Ia** consisting of horizontally and thinly bedded mud with interbedded to interlaminated sand. The sand varies from several grains to several centimetres in thickness. These rhythmites are interpreted to be cyclopels.

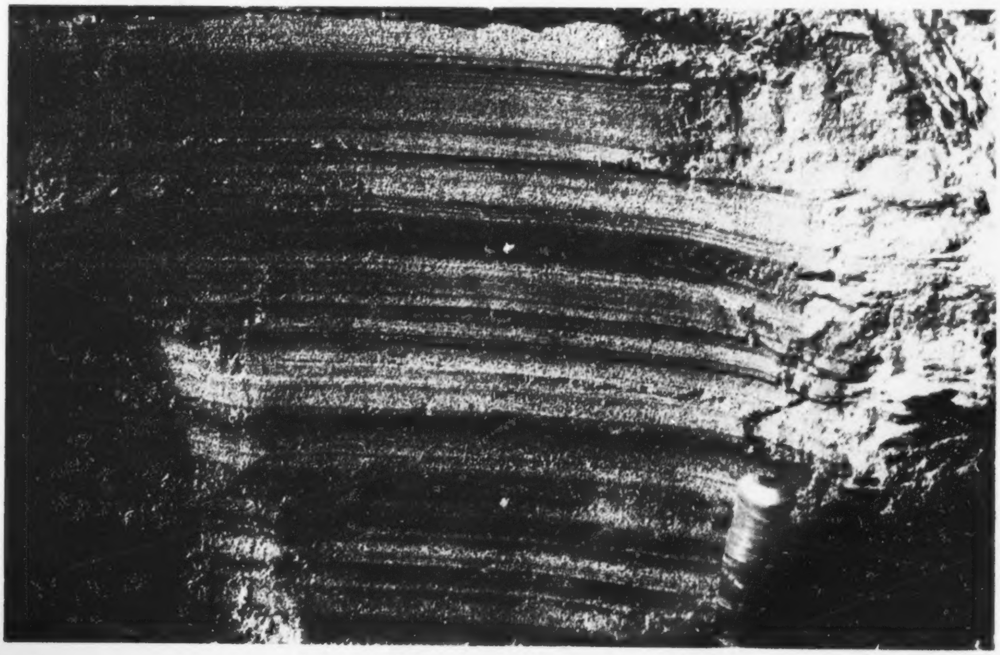


Figure 3.35: Lithofacies **Ib** consisting of horizontal, graded beds of medium to fine sand and silt. They are interpreted to be cyclopsams.

the sediment transported by these plumes. In Alaskan fiords, cyclopsams are deposited within 0.5 km of the ice front whereas cyclopels have been found up to 2.6 km downfiord (Cowan and Powell 1990). Lithofacies II consists of thin to medium graded beds of medium to fine sand, silt and mud (Fig. 3.36). Graded beds have erosional bases with rip-up clasts and dewatering structures. The sand contains parallel laminations and climbing ripples (Fig. 3.36). In some cases, ripples are draped by a thin silt layer. This lithofacies is interpreted to be the product of turbidity currents.

On northernmost Ellesmere Island, Stewart (1991) described similar lithofacies in a fining-upward sequence of deglacial marine sediments. There, subglacial meltwater streams from a rapidly retreating tidewater glacier generated both successive turbidity currents and turbid plumes. Although fining-upward sequences are recognized in the study area, the more typical association is a coarsening-upward sequence from cyclopels to cyclopsams to turbidity current deposits. This facies transition may represent an increasingly more proximal position of the sediment-laden meltwater source which, on western Fosheim Peninsula, is determined by the rate of emergence rather than fluctuations of a tidewater glacier, and/or an increase in discharge during marine regression.

Massive ground ice frequently underlies fine-grained, nearshore deposits (lithofacies Ia) on the peninsula (Fig. 3.37). These ice bodies may have formed by ice segregation during permafrost aggradation shortly after emergence (Pollard 1991) or, they may be buried glacier ice, as proposed for some ground ice bodies in the western Arctic (*cf.* French and Harry 1990). Preliminary observations on the cryostratigraphy and cryogenic texture of the ground ice suggest that ice segregation could be the primary process in its formation (Pollard 1991). However, the age of marine sediments and associated relative sea level adjustments should also be considered in the interpretation of ground ice development.

3.5.2 Marine sediment veneer and washed bedrock

During former marine inundations, weathered bedrock was washed or draped by a discontinuous, marine veneer. In most cases, this veneer is indistinguishable from fine-grained residuum. Isolated clasts resting on bedrock below marine limit may have originated either as dropstones or as the coarse remains of eroded till. Consequently, the differentiation of these two units (Weathered bedrock/Residuum *vs.* Marine sediment veneer and washed bedrock) is arbitrary in nature, based on factors such as elevation, slope, and relationship to adjacent materials.

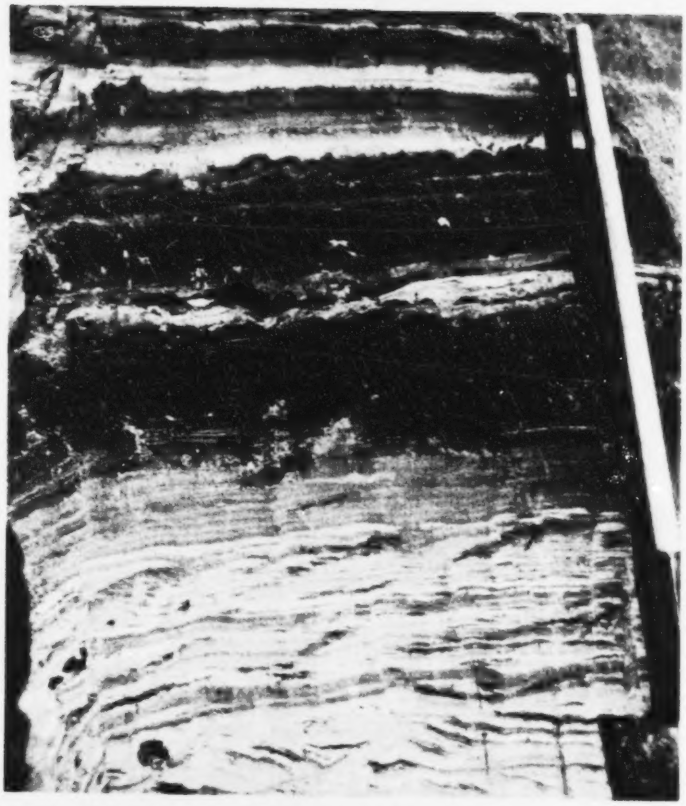


Figure 3.36: Lithofacies II consisting of graded beds of fine sand, silt and clay. Graded beds have sharp erosional bases with rip-up clasts and dewatering structures. The sand contains parallel laminations and climbing ripple structures.

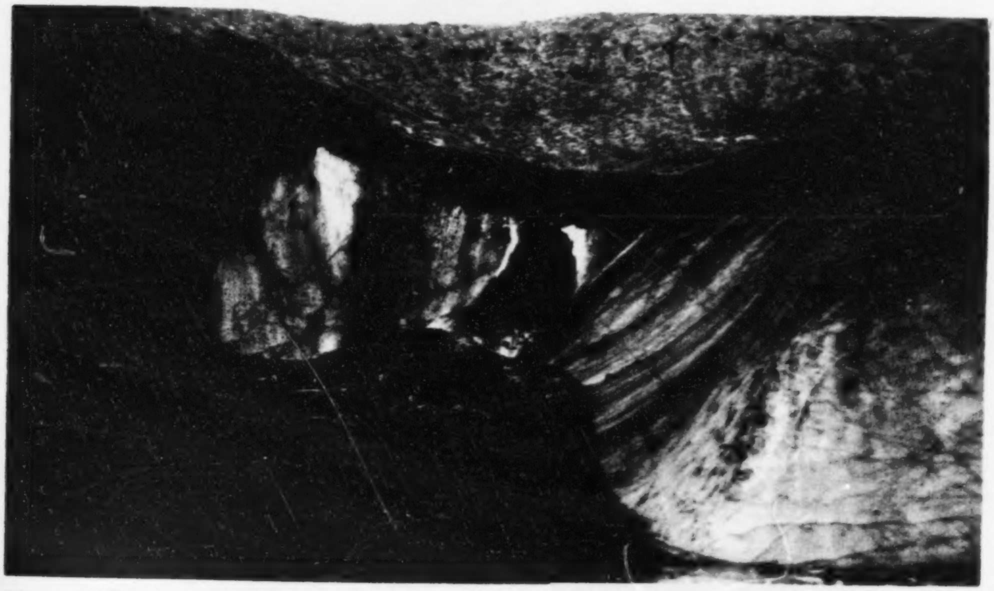


Figure 3.37: Massive ice body (3m thick) underlying lithofacies **Ia** marine sediment (left).

3.5.3 Distribution

Troelsen Ridge

Meltwater from a local ice cap on Troelsen ridge deposited marine sediments in submerged valleys around its perimeter (Fig. 3.1). Above Iceberg Bay lowlands, a small delta at 152 m asl records marine limit (Fig. 3.10). Along Eureka Sound, three valleys drain western Troelsen Ridge. The northernmost valley along Eureka Sound was occupied by an outlet glacier which deposited nearshore sediment up to 94 m asl (Fig. 3.20; site[2] 1). Two separate readings on a washing limit distal to the former ice margin gave elevations of 148 and 152 m asl (Fig. 3.10). The washing limit was indistinct which may account for the difference. Farther south, fossiliferous silt drapes a wave-notched ridge up to 115 m asl (site 2). The third valley along Eureka Sound contains an isolated exposure of fossiliferous silt overlain by deltaic sand and gravel at 70 m asl (site 3). Two washing limits at 134 and 140 m asl provide minimum estimates on marine limit (Fig. 3.10).

Near Eureka weather station, nearshore sediments drape the lower coast and occupy river valleys up to *ca.* 75 m asl. Prominent gravel beaches occur between 110 and 115 m asl west of Station Creek. Farther up Station Creek valley, till at 171 m asl appears washed and small ridges or berms of well-sorted gravel were observed on its surface. These are interpreted as beach berms recording the former presence of a high sea level. The kame moraine in the centre of Weather Station lowlands appears washed to a similar elevation. In summary, the Holocene marine limit in this area is *ca.* 152 m asl but an older sea level reached a maximum of 171 m asl (Fig. 3.10).

Hare Cape ridge

A former outlet glacier draining local ice on Hare Cape ridge terminated in a valley facing Eureka Sound. Downvalley from associated end moraines, a thick wedge of stony marine silt is overlain by beaches up to 106 m asl (Fig. 3.22; site 4). In adjacent valleys, beach berms at 129 m asl represent a minimum estimate on marine limit. North of Blue Man Cape, fossiliferous stony marine silt drapes the local hillslope to 135 m asl and marine limit was recorded by a washing limit at 152 m asl (Fig. 3.10; site 5).

The lowland east of Blue Man Cape contains a thick accumulation of fine-grained marine sediment which originated primarily from meltwater draining the former ice cap on Hare Cape ridge. A sandy delta at 150 m asl marks marine limit and grades downvalley into fossiliferous silt (sites 6 and 7). North and east of Hare Cape ridge, nearshore sediments occupy the lower reaches of river valleys and drape the surrounding lowlands (site 8), including Romulus moraine to within a few metres of its crest (site 9). Washed till marks marine limit at 147-148 m asl (Fig. 3.10).

---

[2] The site number refers to a location on Figure 3.10 where shells were collected for radiocarbon dating. The age of the sample can be found on Table C.1 also using this number; however, the glacial and sea level chronology is discussed in the following chapter.

Black Top Ridge and Snow Goose hill

Meltwater from a local ice cap on Black Top Ridge carried suspended sediment into submerged river valleys and surrounding lowlands. At the southern end of the ridge, an unnamed valley contains nearshore sediment up to 110 m asl (site 10) whereas two nearby washing limits record Holocene marine limit at 145 and 146 m asl (Figs. 3.10 and 3.38). Meanwhile, Hodgson (1985) reported a nearby wave-cut notch at 160 m asl which must relate to a pre-Holocene high sea level.

The northern end of Black Top Ridge is drained by two rivers that flow into Greely Fiord. Although the formerly submerged lowlands appear washed, significant accumulations of marine sediment occured only at much lower elevations (site 14, 60 m asl; site 11, 70 m asl, Fig. 3.39). Two small outwash fans fed by meltwater from the former, upland ice cap record Holocene marine limit at 149 m asl. The delta deposited by regional ice farther east, records a pre-Holocene marine limit at 170 m asl (Fig. 3.10). Gravel terraces associated with the delta were planed at 148 m asl by the later transgression, imparting a step-like appearance.

In Eastwind Lake lowlands, fossiliferous marine silt drapes till to at least 137 m asl, approximately 10 m below the local marine limit (site 13). Nearshore sediments occur discontinuously on the northern slope of Snow Goose hill (site 12, 95 m asl) where marine limit is recorded by beaches (148 m asl) and a washing limit (149 m asl). On the western slope, outwash surfaces have been planed and reworked, forming washing limits at 147 and 148 m asl (Figs. 3.10 and 3.40).

Central Eureka Sound

Five steep valleys drain the southern slope of the Sawtooth Range. Each valley contains evidence of former local ice, either as outlet glaciers from an ice cap or as cirque glaciers occupying presently ice-free basins. During retreat of these glaciers, meltwater deposited small deltas (sites 16-18). The highest occurrence of *in situ* shells in this area (101 m asl; site 15) and a washing limit immediately south of Airfield ridge (139 m asl) represent minimum estimates on marine limit.

Central lowlands

Thick sequences of fine-grained, marine sediment (up to 36 m thick) infill the Slidre River valley (sites 20-22; Figs. 3.41 and 3.42). From this main fill, discrete wedges extend into individual stream valleys draining the Sawtooth Range (Figs. 3.1 and A.7). In contrast, stream valleys on the northern side of Slidre River contain only a thin drape of marine silt (site 23). Deltas are associated with the sediment wedges, recording marine limit between 140 and 143 m asl (Fig. 3.10). The well-sorted, fine-grained nature of this marine sediment contrasts sharply with the adjacent weathered bedrock and till veneer. The upper limit of the sediment wedge varies between 127

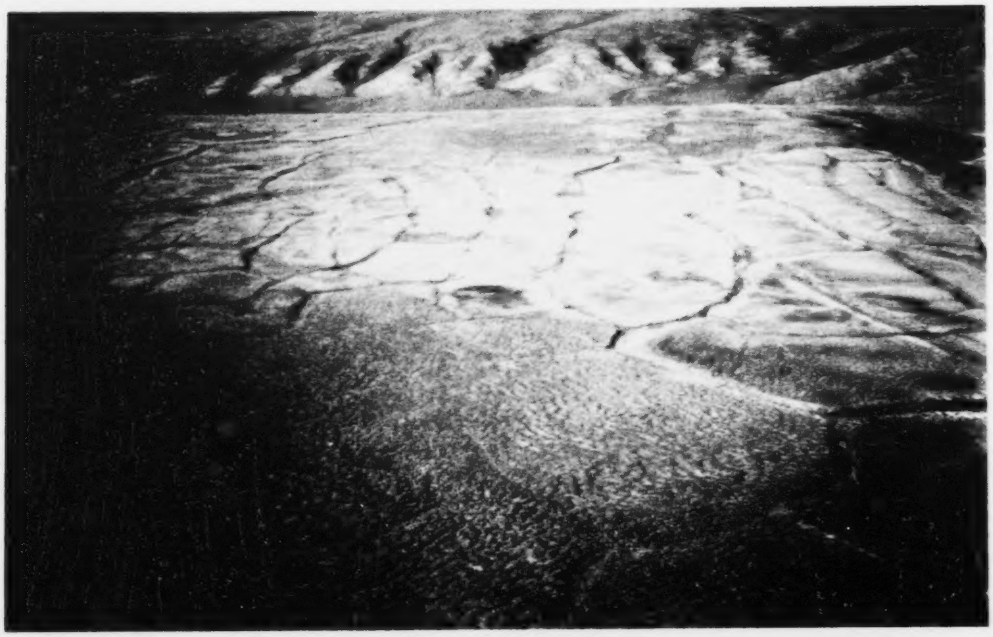


Figure 3.38: Oblique view of marine limit at 145 m asl near Remus Creek. The lighter tone of the washed sandy surface is in marked contrast to the bouldery till veneer upslope (foreground).

Figure 3.39: Outcrop of fine-grained marine sediment below 70 m asl on Greely Fiord coast (site 11).

Figure 3.40: View looking west across planed outwash surface with beach berms (arrowed). The outwash originated from meltwater channels draining Snow Goose hill. Eastwind Lake lowland and Troelsen ridge in background.

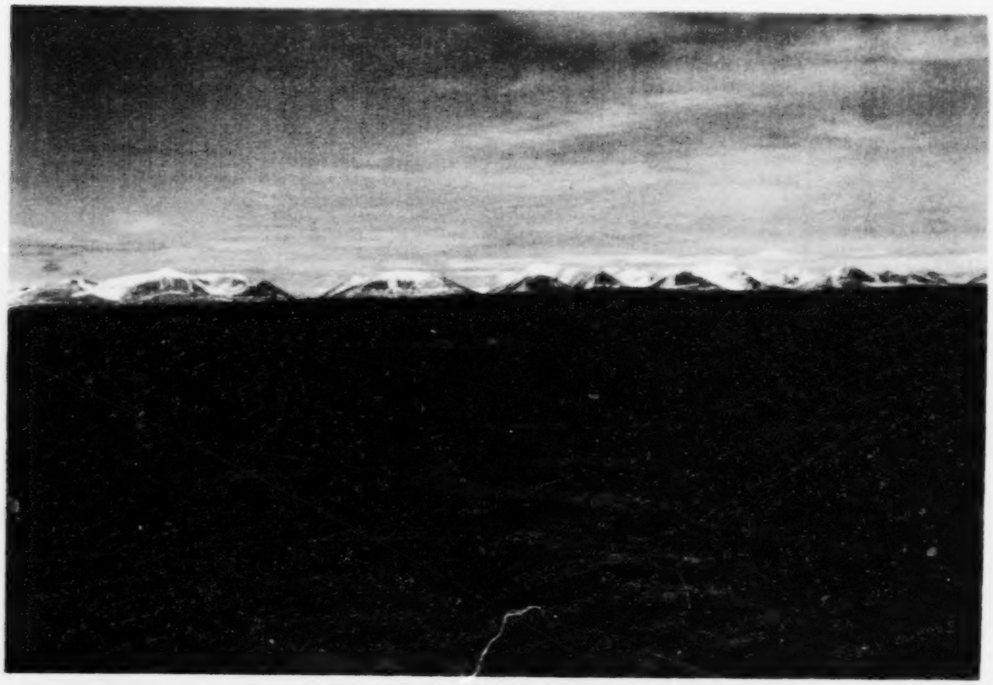


Figure 3.41: View looking southeast across the flat, fine-grained surface of marine sediments in Slidre River Valley. The Sawtooth Range in the background is 35 km distance.

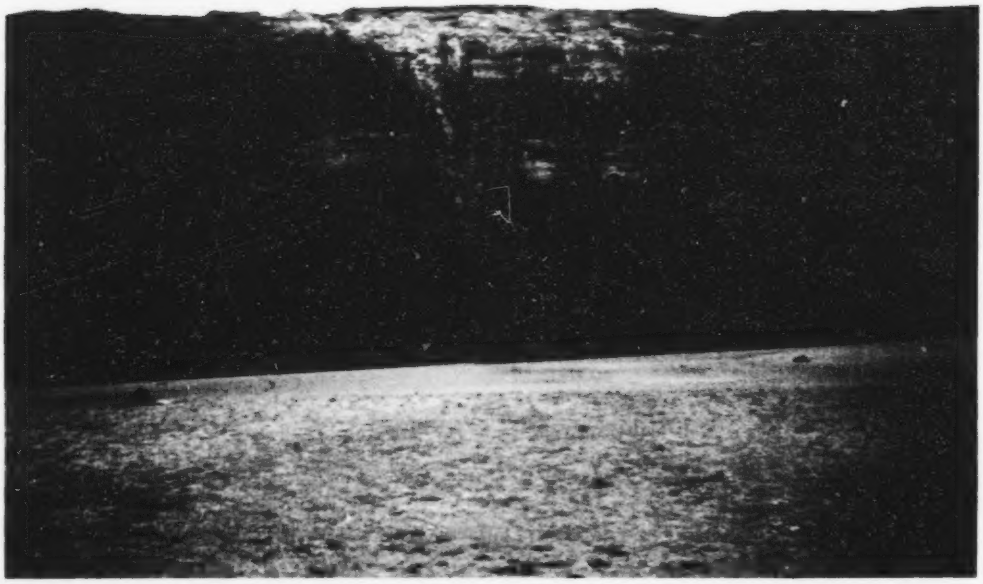


Figure 3.42: A 36 m thick section of marine sediments in Slidre River Valley. The lower half of the section consists of clayey silt rhythmites (lithofacies **Ia**) whereas the upper half is dominated by interbedded sand and silt (lithofacies **Ib** and **II**). The beds dip downvalley at 10-15°.

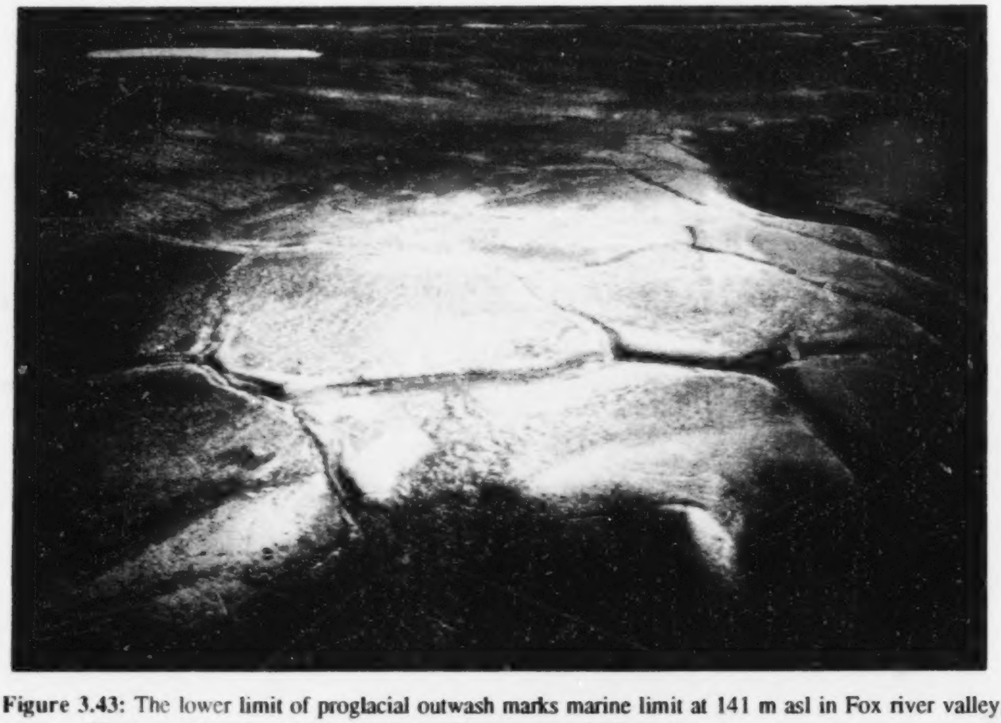


Figure 3.43: The lower limit of proglacial outwash marks marine limit at 141 m asl in Fox river valley. The outwash plain can be traced upvalley for several kilometres.

and 132 m asl and descends to 36 m asl near Slidre Fiord. Considering the minor amount of marine sediment on the western side of Romulus moraine in Slidre River valley, the moraine probably defines the distal margin of a former marine basin that trapped most of the fine-grained sediment (Fig. 3.1).

Delta remnants occupy river valleys draining into Greely Fiord (Fig. A.4). Their elevation is estimated to be 150 m asl based on the 1:50 000 scale topographical map (contour interval 10 m). Meltwater from a former ice cap near the Sawtooth Range drained into marine embayments of Cañon Fiord depositing marine limit deltas at 143 and 144 m asl (Fig. A.3). An outwash terrace in Fox river* valley grades to marine limit at 141 m asl (Fig. 3.43). These elevations are consistent with marine limits reported from elsewhere in Cañon Fiord (Fig. 3.10; Hodgson 1985; England 1990; Sloan 1990).

3.6 FLUVIAL SEDIMENTS

Fluvial gravel and sand comprise a minor surficial unit in the study area. Two types are identified: deposits that lie within the present floodplain and those that are perched above it. The former are flooded every year or two and have channel-scarred surfaces. Inactive alluvium is separated from the floodplain by a distinct bluff and ice-wedge polygons are common on its surface (Fig. 3.43). In some valleys, the highest inactive alluvium consists of proglacial outwash (*e.g.* Fox river valley).

3.7 MINOR SEDIMENTS (not mapped)

3.7.1 Eolian sediment

Wind-blown sand and coarse silt lie adjacent to poorly vegetated, fluvial sediments in the study area. The largest occurrence is on the north side of lower Slidre River floodplain (Fig. A.7). Eolian landforms consist mainly of ripples and low dunes. It has been demonstrated that eolian erosion and deposition are significant processes during winter on Fosheim Peninsula (Edlund and Woo 1992). Although snow cover protects most surfaces, redistribution of snow by wind can create bare patches that become critical sites for eolian activity. In some cases, sand and gravel (including platy clasts with long axis up to 60 mm) have been transported over a kilometre during wind storms when gusts exceeded 90 km/h (Edlund and Woo 1992).

3.7.2 Organic accumulations

Thin organic deposits (peat), composed of mosses and vascular plants, are accumulating in abandoned stream valleys and surface depressions (Edlund *et al.* 1989). Thicker deposits (0.5->5 m) exposed by stream incision are commonly of mid-Holocene age (Hodgson *et al.* 1991). However, this does not necessarily indicate a more favourable paleoclimate at that time but instead, may reflect a bias in the dated samples which have been readily obtained from the Slidre River valley-fill (Hodgson *et al.* 1991). Hence, the dates may be more representative of local emergence than a climatic optimum.

CHAPTER FOUR

Quaternary history

This chapter provides a synthesis of the glacial and sea level history of western Fosheim Peninsula. Information on the surficial geology and geomorphology is integrated with radiocarbon dates, amino acid ratios, marine limits, and stratigraphic sections to reconstruct the Quaternary history. The last glaciation and associated sea level adjustments are discussed first, followed by details and age assignments of older events.

4.1 LAST GLACIATION

Surficial mapping of western Fosheim Peninsula has identified a suite of sediments and landforms that demarcate the extent of the last glaciation (Fig. 4.1). Discontinuous moraines and meltwater channels that either encircle broad uplands or occupy the lower valleys of cirque basins indicate a restricted ice cover. In all cases, these landforms clearly cross-cut weathered bedrock and older glacial deposits. Some glaciers contacted the sea whereas others terminated on broad lowlands above sea level. During deglaciation, sediment-laden meltwater reached the sea depositing deltas at marine limit and thick sequences of glaciomarine sediments in submerged valleys (Fig. 4.1).

4.1.1 Glacial limits

The maximum configuration of the last glaciation is poorly documented because ice-marginal channels are rare; however, cross-valley moraines occupy major drainage systems on Troelsen and Hare Cape ridges, demonstrating that outlet glaciers from local upland ice caps, advanced less than a kilometre downvalley (Fig. 4.1). Along the northwestern margin of Black Top Ridge, outlet glaciers advanced almost 2 km onto the adjacent lowland where they deposited till and discontinuous moraines (Fig.

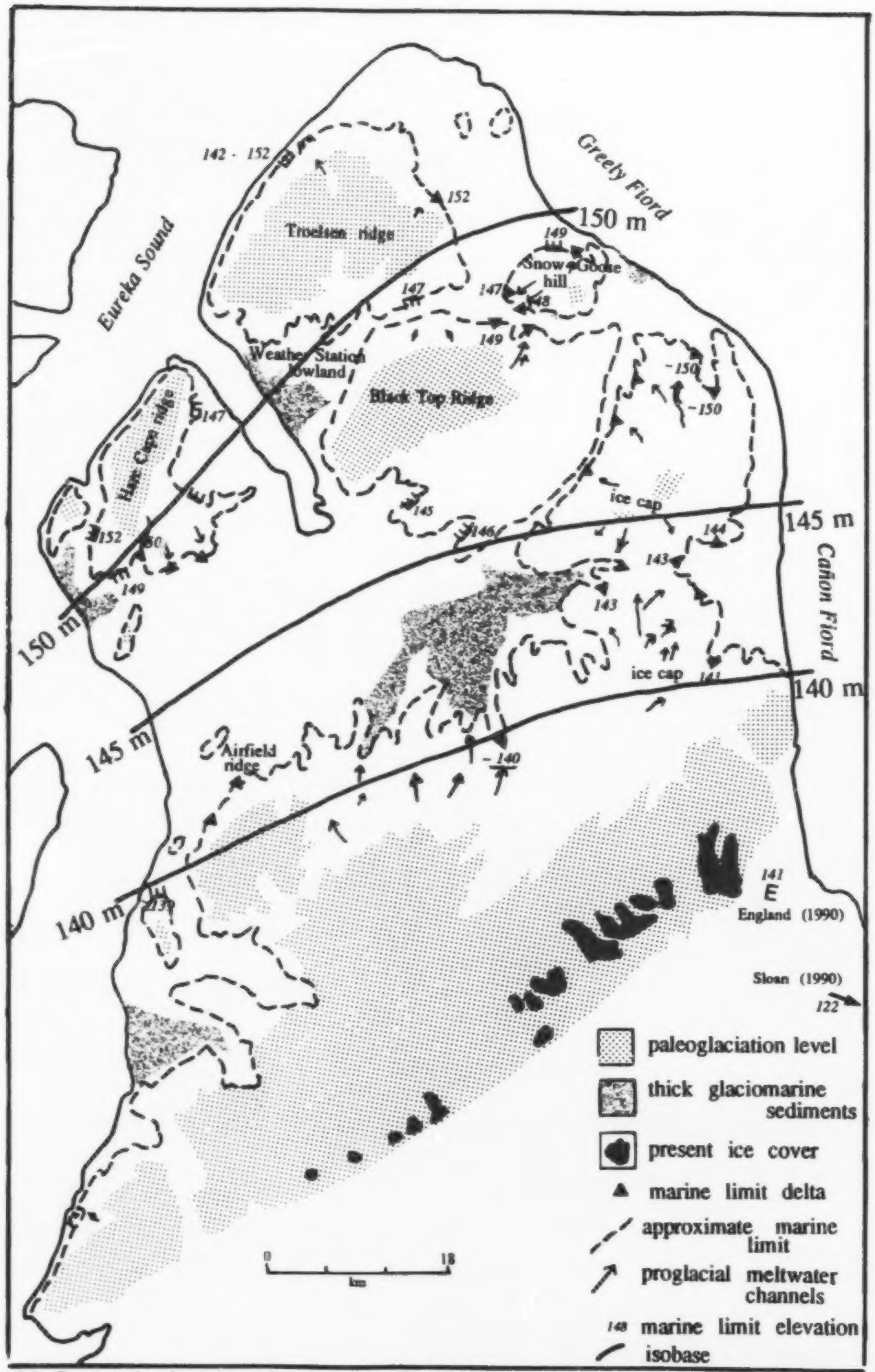

Eureka Sound
Greely Fiord
Cañon Fiord
142 - 152
152
Troelsen ridge
150 m
149
Snow Goose hill
147
148
149
Weather Station lowland
Black Top Ridge
-150
-150
Hare Cape ridge
147
ice cap
145
146
145 m
144
143
152
150
149
150 m
143
ice cap
141
140 m
145 m
Airfield ridge
-140
140 m
139
141
E
England (1990)
Sloan (1990)
122
0
18
km
paleoglaciation level
thick glaciomarine sediments
present ice cover
marine limit delta
approximate marine limit
proglacial meltwater channels
148 marine limit elevation
isobase

Figure 4.1: Proposed paleogeography of western Fosheim Peninsula during the last glaciation. Marine limit elevations included from England (1990) and Sloan (1990). See text for discussion.

4.1). Meltwater from former ice caps on Troelsen, Hare Cape, Black Top, and Airfield ridges and Snow Goose hill deposited deltas at marine limit by at least 8280 BP (Fig. 4.1). The absence of basal flow features on ridge summits suggest that these ice caps may have been frozen to their beds.

Moraines and meltwater channels in the Sawtooth Range demonstrate that modern cirque glaciers were more extensive, glaciers grew in presently unoccupied cirque basins, and ice caps occupied plateaus during the last glaciation (Fig. 4.1). At the southern end of the range, small outlet glaciers advanced onto the Eureka Sound coast where moraines mark their probable maximum limit. In some cases, these glaciers advanced across older fossiliferous till without producing any significant disturbance (Fig. 3.26). Farther north, glacier advances were more extensive. Valleys draining a broad plateau near the Sawtooth Range contain early Holocene glaciomarine sediments, indicating that ice occupied the plateau during the last glaciation. A radial pattern of proglacial meltwater channels, 20-25 km beyond present ice margins in the northern range, demonstrates that a local ice cap grew on the foothills (200-300 m asl), in addition to more extensive cirque glaciers. These channels are eroded in till related to a previous glaciation and are traced to raised deltas, marking Holocene marine limit.

A minor ice cap occupied higher ground on the central lowlands, northeast of Slidre River (Fig. 4.1). Although the former ice configuration is unknown, sediment-laden meltwater reoccupied older meltwater channels and eroded new channels as it drained to the sea. Deltas record Holocene marine limit in these drainages (Figs. 4.1 & A.3).

4.1.2 Marine limit

The magnitude of glacioisostatic depression is an integrated response to regional ice loading and therefore, the marine limit on western Fosheim Peninsula is not only related to the inception of local ice caps but, more importantly, to adjustments in the regional ice cover (*e.g.* Agassiz Ice Cap, Müller Ice Cap). Moreover, the age of marine limit reflects the history of regional ice unloading (and glacioeustatic factors) and may be independent of local ice adjustments. The restricted nature of the last ice cover on western Fosheim Peninsula implies that marine limit records submergence in the peripheral depression (*cf.* Walcott 1970; England 1983). Isobases drawn on marine limit, demonstrate that submergence reached a maximum of *ca.* 150 m asl on northwestern Fosheim Peninsula and declined to the southeast (Fig. 4.1). Both the 140 m and 145 m isobases can be traced southwestward across the central lowlands and southward to central Eureka Sound. The 150 m isobase intersects the peninsula west of Snow Goose hill and runs parallel to northern Eureka Sound through Blue Man Cape. (The elevations of marine limit deltas near Cape Lockwood were derived from topographical maps. The large error associated with this method precludes their use in the construction of isobases.)

The isobase pattern for western Fosheim Peninsula cannot be assigned an age based on this study because dateable material that pertains directly to marine limit was not observed. However, isobases drawn on the limit of the Innuitian Sea in inner Greely and Cañon fiords are consistent with the pattern on western Fosheim Peninsula and span an interval of 7800-8600 BP (England 1992).

4.1.3 Deglaciation

The pattern of ice recession is normally inferred from the distribution of ice marginal features; however, moraines are rare and nested sets of valleyside meltwater channels occur in only a few drainages. It appears that ice caps either melted *in situ* or they retreated rapidly leaving limited geomorphic evidence of their recessional history. Meltwater draining to the sea through presently abandoned or grossly overfit channels built deltas at channel mouths and carried suspended sand, silt and clay into marine embayments (Fig. 4.1). The ages of these sediments provide proxy information on the history of ice recession.

The wide range of dates on glaciomarine sediments (9560-5180 BP, Table C.1) suggests that deglaciation was slow and continued through the mid-Holocene. The timing of initial runoff can be estimated for each ice cap using the oldest date from associated glaciomarine sediments. On western **Troelsen ridge**, fossiliferous stony silt outcrops on the distal side of an end moraine that likely marks the maximum advance of a local outlet glacier. A shell sample from the silt surface dated 8440 ± 80 BP (TO-2233; site 1). At the southern end of **Hare Cape ridge**, fossiliferous silt, containing *in situ* shells dated 8280 ± 80 BP (TO-2238; site 6), occurs downvalley from deltas at marine limit. Meltwater draining from **Black Top Ridge** to Slidre Fiord deposited glaciomarine sediments that contain shells dated 8520 ± 80 BP (GSC-4708; site 10). At the northern end of the ridge, meltwater draining into Greely Fiord deposited a thick sequence of glaciomarine sediments. A *Salix* sp. twig from the base of the sequence dated 44 420 ± 840 BP (TO-2242; site 11) which is too close to the limit of radiocarbon dating to be considered finite. Overlying rhythmites contain *in situ* shells that dated 9080 ± 80 BP (TO-2236; site 11). At the southern end of the **Sawtooth Range**, nearshore sediments associated with meltwater draining from a former ice cap and/or cirque glacier contain shells dated 8430 ± 70 BP (TO-2245; site 15). Farther north, meltwater from local ice in the Sawtooth Range drained into the Slidre River basin, and deposited thick glaciomarine sequences. The oldest shell sample[1] dated 9560 ± 90 (TO-2240; site 19). In summary, initial runoff from the Sawtooth Range had commenced as early as 9500 BP whereas ice caps on local ridge summits were producing meltwater by at least 8280 BP.

---

[1] Echinoderm skeletal remains (brittlestar?; A. Aitken, pers. comm. 1991), from 2.5 m below sample TO-2240, dated 8900 ± 90 BP (TO-2611), and although the outer 48% of the sample was leached, the porous nature of brittlestar spicules renders them more susceptible to contamination than shell material.

Remnants of raised marine deltas occupy valleys that drained former ice caps and cirque glaciers in the southern Sawtooth Range (Fig. 4.1). Bottomsets underlying the northernmost delta contain shells dated 6480 ± 70 BP (TO-2244; site 16) whereas shells from foresets dated 5180 ± 90 BP (GSC-5147; site 17). Foresets of a nearby higher delta contain shells dated 6330 ± 90 BP (GSC-5117; site 18). The coarse nature of the delta foresets suggests that ice was still present in these drainages at that time. At site 11, near Greely Fiord, glaciomarine sediments were accumulating at a rate of 60 cm/century between 8000 and 9000 BP (Table C.1). In Slidre River valley, the sedimentation rate was ≥64 cm/century[2] between 5900 and 4000 BP (site 22), demonstrating that deglacial runoff and high glaciomarine sedimentation continued to this time. Therefore, it appears that local ice caps were maintained into the mid-Holocene and that deglaciation progressed slowly.

4.1.4 Postglacial emergence

The postglacial emergence of western Fosheim Peninsula cannot be reconstructed based on the radiocarbon dates obtained in this study because the contemporary sea level for most dated samples is unknown. Instead, the elevation and age of samples are plotted on Figure 4.2.. Because the shell species used in this study (Fig. C.1) live at various water depths, their contemporary sea level, and hence, the postglacial emergence curve must lie between the sample elevation and the marine limit. Published curves from adjacent areas are discussed in the next section (Fig. 4.2; Hodgson *et al.* 1991; England 1992).

The highest shell sample dated 8990 ± 80 BP (TO-2232; site 13). It was collected from a stony silt veneer at 137 m asl on Eastwind Lake lowlands, 10-12 m below the marine limit. Near Blue Man Cape, fossiliferous marine silt drapes the local hillside to 135 m asl[3], or 17 m below marine limit. A shell sample from the silt surface at 132 m asl dated 8680 ± 80 BP and 8640 ± 80 BP (GSC-5156 & TO-2237; site 5). Fifteen shell samples, collected at least 35 m below the local marine limit, dated between *ca.* 7900 and 8500 BP. GSC-4957 from 110 m asl on Romulus lowlands (site 9) dated 8230 ± 80 BP and is the youngest and highest sample of the group. Two radiocarbon dates reported by Hodgson *et al.* (1991) are also plotted on Figure

---

[2] This sedimentation rate is calculated using radiocarbon dates on shell (5870 ± 80 TO-1965, 21 m asl, Table C.1) and peat (3970 ± 80 GSC-2039, 35 m asl, Hodgson *et al.* 1991). The shell sample was overlain by 12 m of lithofacies **Ia** and 3 m of lithofacies **Ib** and **II** glaciomarine sediment. This sediment must have been deposited by 4000 BP because sea level was below the top of the section at this time as demonstrated by local peat accumulation at 35 m asl. Therefore, dates from the shell and the peat encompass the marine sedimentation period. A minimum average sedimentation rate is calculated for lithofacies **Ia** only.

[3] J.G. Fyles collected a shell sample dated 8710 ± 140 BP (GSC-254) at a reported elevation of 140 m asl in this vicinity (Dyck *et al.* 1965). However, repeated measurements on the limit of fossiliferous marine silt provided a maximum elevation of only 135 m asl.

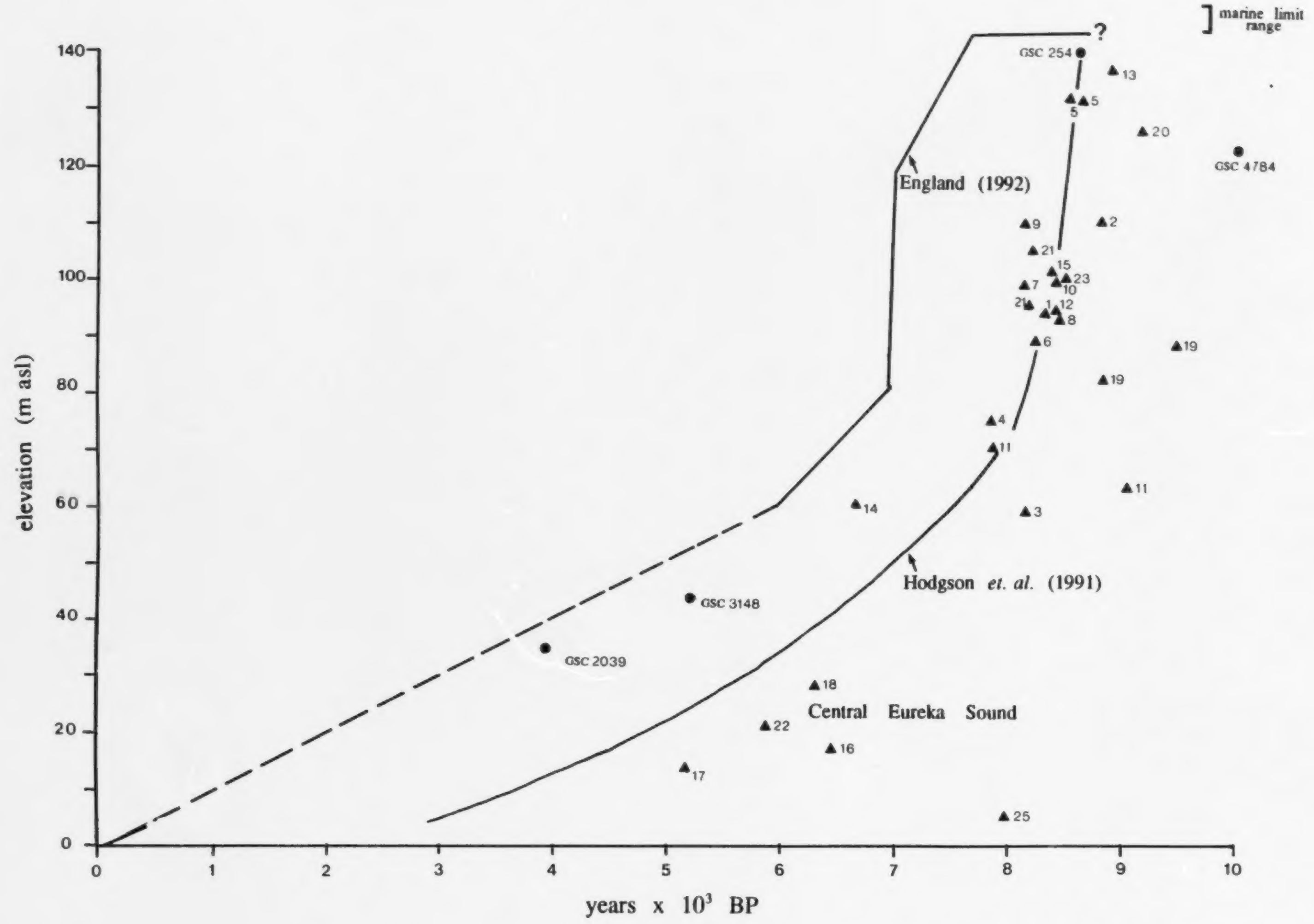


Figure 4.2: Minimum emergence curve for northern Fosheim Peninsula. Numbers refer to shell sample sites (see Table C.1 and Fig. 3.10). Emergence curves from inner Greely Fiord (England 1992) and Fosheim Peninsula (Hodgson *et al.* 1991) are also shown.

4.2. A peat horizon at 35 m asl in Slidre River valley dated 3970 ± 80 BP (GSC-2039), indicating that sea level was <35 m at 4000 BP. A *Salix* twig recovered from delta-front deposits, also in Slidre River valley, dated 5280 ± 60 BP (GSC-3148). The age of the twig represents a maximum date for a sea level ≥44 m asl (Hodgson *et al.* 1991).

Three shell samples from deltas at the southern end of the Sawtooth Range provide estimates on sea level between 5200 BP and 6300 BP in this area. The highest sample, dated 6330 ± 90 BP (GSC-5117; site 18), relates to a former sea level recorded by topsets at 30 m asl. Shell sample TO-2230 demonstrates that sea level was at least 30 m higher in Greely Fiord at this time. Consequently, the 6500 BP shoreline has a delevelled gradient of at least 0.3 m/km, sloping towards the south or southeast. Shell sample GSC-5147 from foreset beds in a nearby delta dated 5180 ± 90 BP (site 17) and relates to a former sea level at 16 m asl (elevation of topset/foreset contact).

4.1.5 Discussion and regional correlation

The extent, configuration and age of the last glaciation on western Fosheim Peninsula (*research objective #1*) are reconstructed using geomorphic, sedimentologic, and chronologic evidence. Moraines and meltwater channels outline the broad configuration of the last ice cover which was generally restricted to small ice caps on ridge summits and cirque glaciers in the Sawtooth Range (Fig. 4.1). These landforms cross-cut weathered bedrock and older glacial deposits. The distribution of glaciomarine sediments is confined to formerly submerged drainages that had local ice in their upper reaches. These sediments consist of fine-grained rhythmites derived from turbid overflow plumes, overlain by coarser-grained sediments from turbidity currents. This facies transition may represent either an increasingly more proximal position of the meltwater source during marine regression or a marked increase in sediment supply causing extensive progradation of coarse sediments. This sequence differs from the deglacial sedimentary record associated with retreating tidewater glaciers on northern Ellesmere Island, where overstep and fining-upward sequences have been reported (Stewart 1991). The chronology associated with glaciomarine deposits supports a restricted ice extent, indicating that initial sedimentation occurred in Slidre River valley at the centre of the peninsula.

Ice extent

The combined evidence for the last glaciation suggests that western Fosheim Peninsula was not inundated by the Innuitian Ice Sheet at this time (*research objective #1)* and that it lay beyond the limit of the larger ice caps (*e.g.* Agassiz Ice Cap, Central Ellesmere Ice Cap). These conclusions are in agreement with the limit of the

Franklin Ice Complex in inner Greely and Cañon fiords, where outlet glaciers from the Agassiz Ice Cap advanced only 5-60 km (England 1990, 1992; Sloan 1990). Where these glaciers reached the fiords, the sea constrained farther advance; consequently, the fiords were occupied by the Innuitian Sea, rather than ice, during the last glaciation (England 1983, 1990, 1992; Sloan 1990).

Initiation and buildup of the last ice cover remains poorly dated (England 1990; Lemmen 1989). England (1992) presented two hypotheses which represent end members for ice buildup to the last glacial maximum. His first hypothesis had glaciers reach their maximum before 18 000 BP; however, this requires that ice buildup occurred during a period (70-14 ka BP) when precipitation in the region was only a third of present values (Firestone *et al.* 1990). His preferred hypothesis involves a rapid buildup of ice after 14 000 BP in response to increased precipitation. Given that the climate was colder and drier for most of the last glaciation (>14 000 BP), then aridity was a primary climatic constraint on the magnitude and timing of the last ice maximum in the high arctic (England and Bradley 1978; Miller and de Vernal 1992). Dates on redeposited organics from one locality on northern Ellesmere Island support late ice buildup, suggesting that the last ice limit there was attained as late as 11 000 BP (Lemmen 1990).

The paleoglaciation level on western Fosheim Peninsula during the last glaciation is estimated to be 200-300 m asl, or 700-800 m lower than the present glaciation level. This drop in paleoglaciation level is double the regional average (England 1986; Sloan 1990) and may reflect local climatic factors. In Chapter 2, it was suggested that Eureka Sound lowlands probably received less precipitation than adjacent areas during the last glaciation, based on the present distribution pattern. However, the increased warming of *ca.* 10°C that occurred between 14 and 10 ka BP (Summit, Greenland, Firestone *et al.* 1990) and *ca.* 5°C during the early Holocene on the Agassiz Ice Cap (Koerner and Fisher 1990) may have led to reduced sea ice around the peninsula, providing a local moisture source. Hein and Mudie (1991) concluded that shelf sediments off western Axel Heiberg Island represent relatively open water by at least 9950 BP. Today, along the northwest coast of Ellesmere Island, glaciation levels decline from 1100 m asl over the Grant Land Mountains to sea level at the coast. This abrupt decline is regarded as a function of local climatic conditions where periodic open water, resulting from leads in the pack ice, causes increased moisture and reduced ablation (Sagar 1962). Similar conditions probably existed in the study area during the last glacial maximum.

Marine limit

Beyond the last ice limit on western Fosheim Peninsula, there is evidence that the Innuitian Sea occupied the peripheral depression. The distribution of marine limit deltas and washing limits are used to draw isobases on this shoreline (Fig. 4.1). The isobase pattern represents an extension of England's (1992) isobases from Greely Fiord (Fig. 1.9) and demonstrates that the northeast-southwest ridge of high emergence diverges in the direction of Eureka Sound (*research objective #2).* This divergence indicates that the ridge is rising towards the west and suggests that marine limit is likely >150 m asl on the eastern coast of Axel Heiberg Island. Hodgson (1985) reported that marine limit is only 100 m in central Eureka Sound (*e.g.* Bay Fiord; Figure 2.1). If so, the 140, 145, and 150 m isobases must converge southwest of Fosheim Peninsula, indicating a plunging ridge in that direction.

It is generally accepted that delevelled shorelines in glaciated regions rise toward centres of maximum former ice thickness (*cf.* Andrews 1970). However, the isobase pattern drawn on the limit of the Innuitian Sea outlines a ridge of emergence that is centred over northern Fosheim Peninsula where the smallest ice load change occurred during the last glaciation. Hence, the pattern is difficult to reconcile with a solely glacioisostatic explanation of postglacial uplift (England 1992). Recent geophysical modelling of the relative sea level history of the high arctic has attempted to explain the emergence pattern in terms of large ice load changes, in contradiction to the geological evidence (Tushingham 1991; England *et al.* 1991). However, when the model input was revised to reflect the geological data, the elevation difference between the observed marine limit and that predicted by a small ice load change was relatively small (England *et al.* 1991). For instance, the marine limit in Bay Fiord (91 m, Hodgson 1985), the nearest site to the study area, was actually *overestimated* by 9 m using the small load change. It is still uncertain whether the anomalously high emergence in the study area can be explained by small changes in ice load and therefore, other factors (*e.g.* the influence of combined peripheral depressions from surrounding ice caps, *cf.* England 1976), which are not fully understood, may be amplifying the glacioisostatic effect.

Alternatively, the known history of rifting and the presence of seismically active areas in the high arctic suggest that tectonic forces may also have contributed to Holocene emergence (Hodgson 1981; England 1992; Dyke *et al.* 1991). England (1992) recognized that the isobase pattern in Greely Fiord coincides with the orientation of the local geologic structure (*i.e.* the boundary between the Sverdrup Basin and the Franklinian Mobile Belt; Fig. 1.3), and hence, suggested that the pattern may be attributable in part to Holocene tectonic uplift or structurally-controlled glacioisostatic uplift. With the extension and refinement of the isobase pattern across western Fosheim Peninsula, it may be important to compare the orientation of the ridge with the boundaries of the Fosheim Basin. This was one of several sub-basins that evolved from the Sverdrup Basin following the earliest phase of Eurekan tectonism in the Late Cretaceous. Details on post-Eurekan tectonic activity in the high arctic remain limited, although it is recognized that Fosheim Basin was partially rejuvenated during the Miocene and/or Pliocene when high terrace sediments were deposited across west-central Ellesmere Island and eastern Axel Heiberg Island (Miall 1986; discussed later). Although Dyke *et al.* (1991) have demonstrated some structural control in the postglacial isobase pattern of the central arctic, the results of this study lack sufficient resolution to discriminate glacioisostatic and possible neotectonic components in the isobase pattern of the Innuitian Sea (*research objective #2*).

England (1992) described a period of high sea level in Greely and Cañon fiords between 8800 and 7800 BP, when slow postglacial uplift equalled the rate of eustatic rise, producing a stable relative sea level at marine limit (Fig. 4.2). Although the postglacial emergence for the study area cannot be resolved at present, sea level was within 20 m of marine limit at 10 600 BP and within 10 m at 9000 BP (Fig. 4.2). This prolonged period of high sea level on western Fosheim Peninsula (at least 10 600-9000 BP) suggests that the balance between unloading and eustatic rise may have begun as early as 10 600 BP; however, the duration of this relative sea level stability is unknown.

Deglaciation

The study area contains the oldest dated sediments associated with initial runoff from the last ice cover on west-central Ellesmere Island (9560 BP[4], site 19). Here, the melt period begins the earliest, lasts the longest, and has the highest number of snow-free days (Edlund and Alt 1989). Although there is wide variability in dates (9560-8280 BP), the earliest runoff is associated with ice caps and glaciers in the Sawtooth Range, followed by those along the coastal margins. This general chronology corresponds to initial runoff dates from adjacent areas (Bay Fiord, 9140 ± 310, GSC-3388, Hodgson 1985; outer Cañon Fiord, 8850 ± 60 BP, TO-1285, England 1990) and to the summer melt record in ice cores from the Agassiz Ice Cap (9500-8500 BP; Koerner and Fisher 1990).

Local ice caps in the study area were maintained through the mid Holocene (5000-6000 BP) as demonstrated by the high glaciomarine sedimentation rates in formerly submerged basins (sites 16, 17, 18 and 22). In contrast, outlet glaciers from larger ice caps (*e.g.* Agassiz and Central Ellesmere ice caps) retreated behind their present margins by 7500 BP (Hodgson 1985; England 1990; Sloan 1990). However, England (1990) also reported very late ice in Borup Fiord where deglacial marine limits are anomalously low. The slow demise of smaller ice caps is attributed to the same local climatic conditions that produced a marked decline in the paleoglaciation level. High arctic ice core records show a continuously warm period from 10 000 to about 4000 BP with the warmest period on the Agassiz Ice Cap between 9000 and 7000 BP (Koerner *et al.* 1988; Koerner and Fisher 1990). Bradley (1990) concluded that summer temperatures during the early to mid-Holocene were probably as high, or higher than they have been for much of this century. This increased warming would have resulted in a reduced sea ice cover which may explain the presence of a narwhal in Greely Fiord at 7970 ± 90 (TO-1966, site 25). Like all marine mammals, narwhal require open water to survive and their migration routes are directly related to seasonal sea ice conditions in channels (Newman 1978). In summary, it is proposed that higher temperatures in the early to mid-Holocene led to increased precipitation and reduced melting which were critical to the continued maintenance of local ice caps on the peninsula (*research objective #3*).

---

[4] The shell sample dated 10 600 ± 90 (GSC-4784; Hodgson *et al.* 1991) was collected from the surface of a marine silt veneer and therefore, its relationship to glacial runoff cannot be determined.

Postglacial emergence

The postglacial emergence history of western Fosheim Peninsula cannot be resolved from the present data set (Table C.1). However, these data demonstrate that the provisional curve of Hodgson *et al.* (1991) should be modified because younger samples plot above their curve (Fig. 4.2). England's (1992) emergence curve from inner Greely Fiord represents a maximum reconstruction for data from western Fosheim Peninsula because it assumes that all samples older than 7800 BP relate to a former sea level at marine limit (Fig. 4.2). If this maximum interpretation is correct, then the apparent absence of shells between 7000 and 8500 BP at 'higher' elevations on western Fosheim Peninsula must have local paleoenvironmental significance because England (1992) reported many samples from this age range within 10-15 m of marine limit in his study area. Perhaps heavy sea ice and/or highly turbid water constrained the activity of marine organisms in shallow water, close to marine limit (A. Aitken, pers. comm. to J. England 1992).

4.2 PREVIOUS GLACIATIONS

Beyond the last glacial limit on western Fosheim Peninsula, there is abundant evidence of older, more extensive glaciations. Erratics are widely distributed and although they need not represent a single glacial event, high-elevation erratics could have been deposited only by ice that covered the entire peninsula. Erratic dispersal patterns document former ice flow direction whereas amino acid ratios from ice-transported shells differentiate discrete glaciations. The pattern of ice retreat is inferred from the distribution of moraines, meltwater channels and glaciofluvial deposits whereas the occurrence of marine sediments and landforms above the Holocene marine limit suggests that glacial recession was associated with a higher relative sea level.

4.2.1 Ice flow patterns and glaciations

Regional till on western Fosheim Peninsula has been subdivided on the basis of its lithological composition. Till containing granite erratics forms a long, narrow dispersal train on the western fringe of the peninsula, roughly coincident with northern Eureka Sound (Fig. 3.16). It represents a regional ice source. Other erratics, but mainly quartzite, are widespread across the peninsula, including the highest summits, and represent a westerly ice flow from central Ellesmere Island.

Tentatively, the granite dispersal pattern in northern Eureka Sound may be extended southward to Norwegian Bay and westward through Bay Fiord to the Prince of Wales Mountains, based on earlier observations of granite erratics (Fyles 1962; Tozer 1965; Hodgson 1985). The simplest dispersal mechanism that explains the granite distribution involves a westerly ice flow through Bay Fiord from a former centre over the Canadian Shield of southeastern Ellesmere Island (Fig. 1.2). Once in Eureka Sound lowlands, the ice flow probably diverged north and south. To the north, ice overtopped the eastern highlands of Axel Heiberg Island and advanced along northern Eureka Sound to Greely Fiord.

Alternatively, granite erratics may have been transported by the Laurentide Ice Sheet from the Canadian Shield of the lower arctic. The occurrence of crystalline erratics on islands in the western arctic prompted Hodgson (1989, 1990) to speculate that they were deposited by north-flowing Laurentide ice during an extensive pre-Wisconsinan glaciation. However, it is less likely that Laurentide ice advanced through Eureka Sound to Fosheim Peninsula, more than 600 km beyond its Wisconsinan limit, unhindered by Ellesmere ice. Northwest Greenland is precluded as a source for the granite erratics on western Ellesmere Island because the maximum advance of Greenland ice only abutted the east coast of the island (J. England, pers. comm. 1992). Therefore, it is emphasized that a local Ellesmere Island source for granite erratics is the more reasonable hypothesis.

Amino acid analysis was carried out on ice-transported shells from regional and local tills to determine whether multiple glaciations were responsible for the observed erratic distributions. Amino acid ratios, specifically the ratio of alloisoleucine to isoleucine, demonstrate that two shell populations, with mutually exclusive ratios, can be identified according to till location (Tables C.2 and 4.1; Figs. 4.3 and 4.4). The highest ratios (Group I; *free* 0.372 - 0.741; *total* 0.089 - 0.185), and consequently, the oldest shells, are associated with till on the northwestern and southwestern corners of the peninsula. Lower ratios (Group II; *free* 0.146 - 0.355; *total* 0.047 - 0.086) are confined to till on Hare Cape ridge, Romulus lowland and west-central Eureka Sound. Samples with mixed populations (Group III; *free* 0.285 - 0.417; *total* 0.041 - 0.09) occur near Cape Lockwood on the Greely Fiord coast and at the southern margin of the Sawtooth Range in Eureka Sound. The occurrence of two tills with ice-transported shells of discrete relative ages can only be reasonably explained in terms

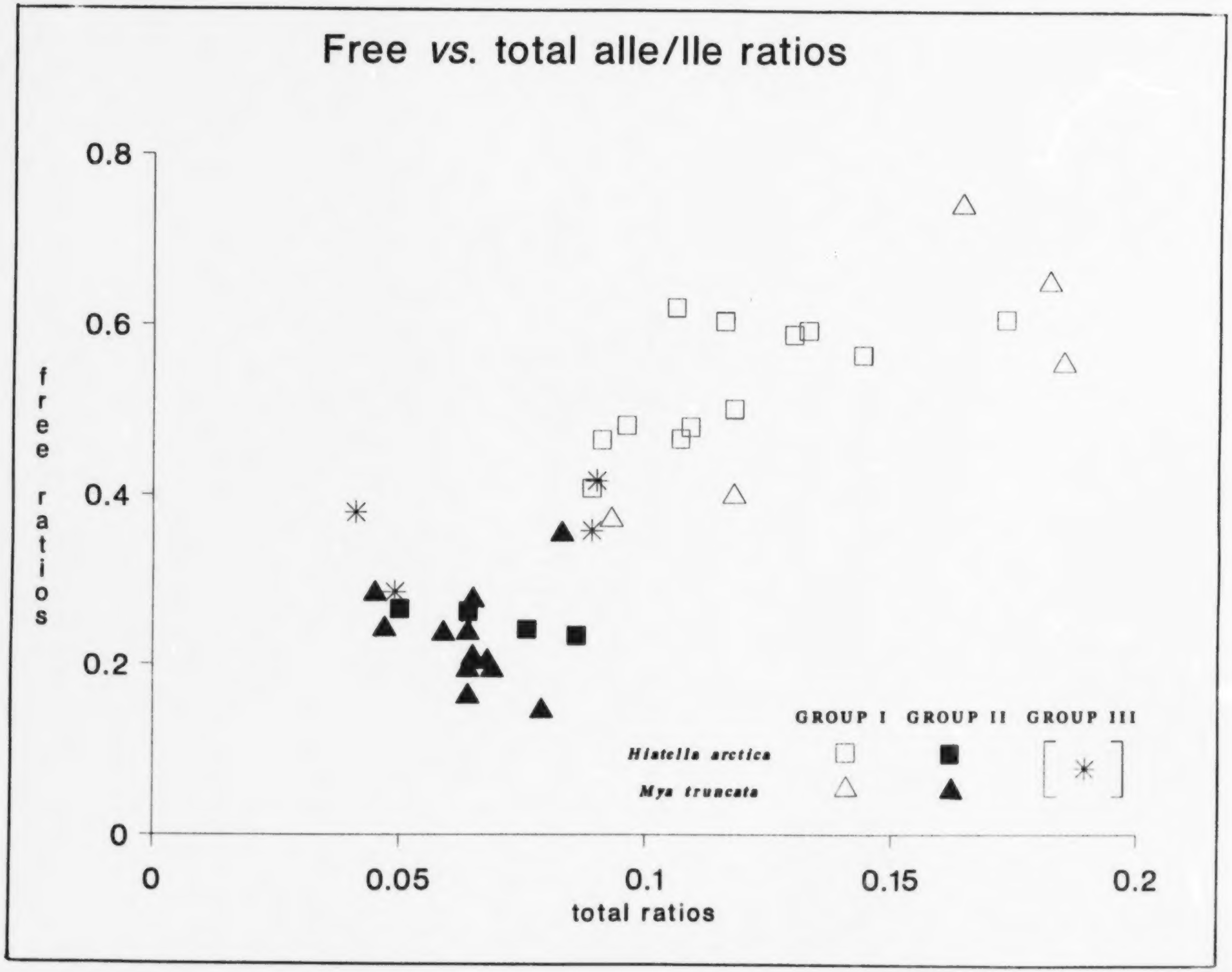


Figure 4.3: Plot of free vs. total alle/Ile ratios from shells in the study area. Samples are classified according to taxa and group membership.

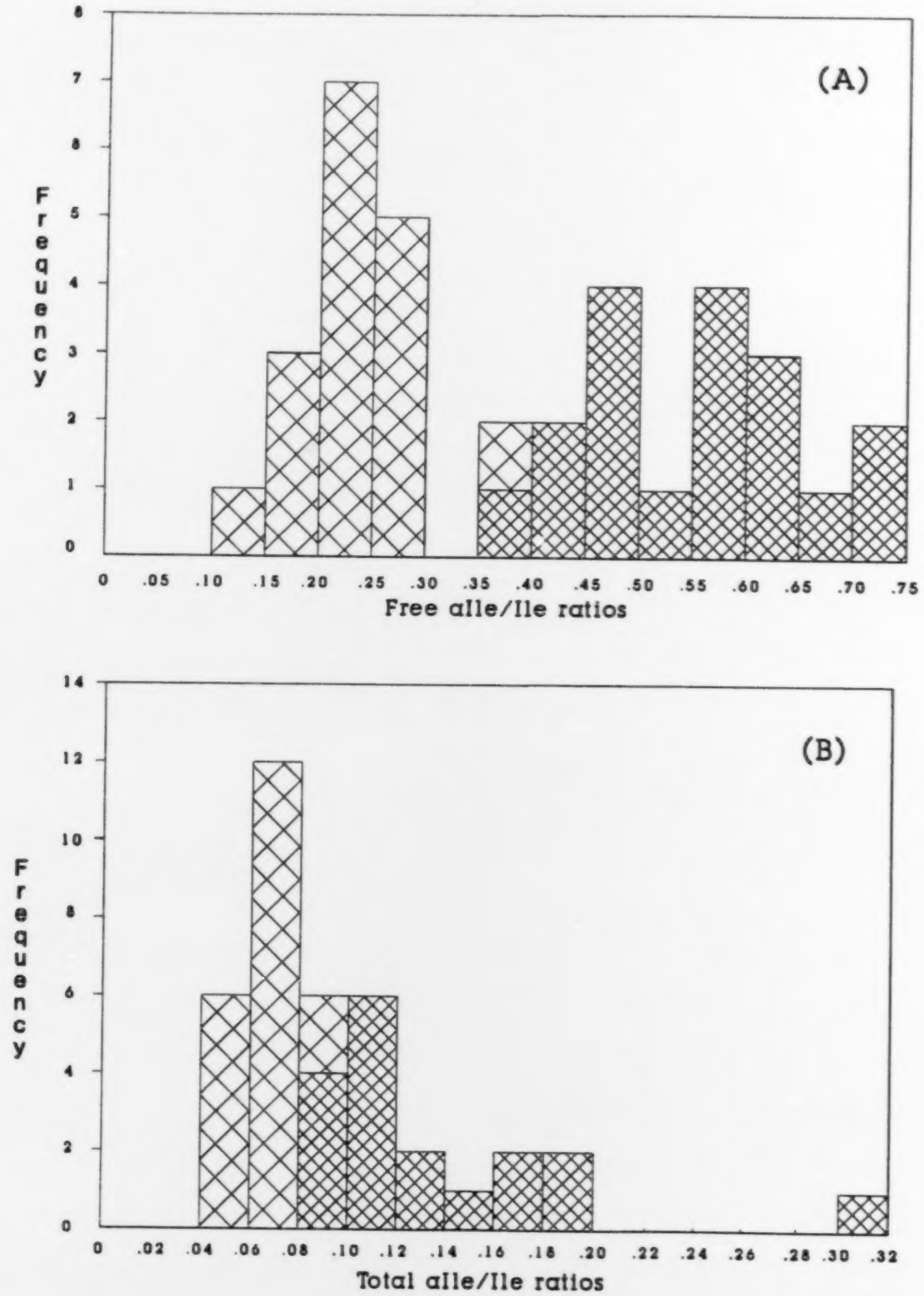


Figure 4.4: Frequency distributions of (A) free and (B) total alle /Ile ratios. Small and large cross-hatch denote Group I and Group II, respectively.

of at least two glaciations. These are provisionally named **Group I** and **Group II** glaciations for the oldest and youngest events, respectively.

Table 4.1: Summary group statistics for aIle/Ile ratios[5] from the study area

	GROUP I		GROUP II		GROUP III	
	Free	Total	Free	Total	Free	Total
range	0.372 - 0.741	0.089 - 0.185	0.146 - 0.355	0.047 - 0.086	0.285 - 0.417	0.041 - 0.090
$\bar{x}$	0.547	0.127	0.232	0.065	*n.a.*	*n.a.*
S_x	0.102	0.032	0.049	0.011	*n.a.*	*n.a.*
CV	19%	25%	21%	17	*n.a.*	*n.a.*
n	19	17	17	20	4	4

n.a. = not applicable
CV = coefficient of variation ($S_x/\bar{x}$)

Amino acid epimerization rates are extremely sensitive to temperature and therefore, it may be argued that all the shell samples in the study area have identical ages but some have experienced different post-depositional thermal histories. Considering the close proximity of the two populations, the only conditions that could act independently in adjacent areas to alter the diagenetic temperature of samples are glacial ice cover and marine submergence. It can be reasonably assumed that under both conditions the mean annual temperature would be close to 0°C (Miller and Brigham-Grette 1989; Sejrup 1990; Lemmen and England 1992) whereas the mean annual air temperature (MAT) at Eureka is -20°C. Such a difference in thermal history would have a significant effect on the rate of epimerization and the amino acid ratios of a sample (Sejrup 1990; Lemmen and England 1992).

If it is assumed that all the shell samples were transported by a single glaciation, then any subsequent regional ice cover would likely affect all sites equally. Alternatively, a local ice cover, similar to the last glaciation, would be restricted to ridges and plateaus, leaving lowlands ice-free. All Group I shell samples were collected from beyond the limit of the last glaciation whereas some Group II shell samples may have been ice-covered on Hare Cape ridge (sites 33-35) and central Eureka Sound (site 40; compare Figs. 4.1 and 3.10). Upland ice caps were considered to be cold-based during the last glaciation and therefore, they would have had a basal temperature between the pressure melt point and the prevailing air temperature. Consequently, the differences in thermal history between ice-covered and ice-free sites during the last glaciation may not have been significant and any minor effect would only have decreased the apparent age difference between the two shell populations.

---

[5] Some amino acid ratios were not used in data reduction or analysis because of (1) taxonomic differences, (2) anomalously high ratios from 'old' shells, or (3) inconsistent free and total ratios suggesting analytical error. See Table C.2 for omitted data.

The location of a shell sample relative to successive marine transgressions may have a cumulative effect on its amino acid ratios (Sejrup 1990; Lemmen and England 1992). Consequently, it is assumed that shells transported above the marine limit would provide lower amino acid ratios than shells of the same age that have experienced multiple transgressions below marine limit. There are at least two marine transgressions (*ca.* 170 m and 150 m asl) of unknown duration recognized on western Fosheim Peninsula. If shell samples are stratified on the basis of elevation relative to the limits of these transgressions (see endnote *e*, Table C.2; Lemmen and England 1992), then the two shell populations are still clearly identifiable *within* each thermal category (Table C.2). Moreover, shell samples from a small area that have experienced different thermal histories (*e.g.* above and below marine limit on Hare Cape ridge) show only a slight increase *between* thermal categories (compare sites 33-36; Table C.2). Such a small difference between categories would be expected if marine transgressions were of short duration, and if sample sites are close to marine limit.

In summary, it is proposed that groups I and II samples represent discrete shell populations deposited during separate glaciations. The oldest (Group I) glaciation was the most extensive, advancing across Iceberg Point lowlands and into Greely Fiord (Fig. 4.5). The youngest (Group II) glaciation was marginally less extensive, terminating somewhere between Hare Cape ridge and Cape Lockwood (Fig. 4.6).

4.2.2 Group I glaciation

Remnants of the older fossiliferous till are found between Iceberg Point lowlands and Black Top Ridge and along central Eureka Sound (Fig. 4.5). Although ice clearly inundated the peninsula at some time, as evidenced by the high-elevation erratics, it is uncertain whether this maximum stage is associated with the oldest recorded till. However, as only a single till was observed on Iceberg Point lowland (sections 2 and 3, Appendix B), this correlation is assumed here.

Non-granite till represents ice movement from the east or southeast whereas granite-rich till demonstrates that ice flowed north from central Eureka Sound. Although the profile of granite erratics on eastern Troelsen ridge cross-cuts non-granite till (Fig. 3.16), amino acid ratios on shells from the two tills are indistinguishable (compare sites 26-28 and sites 29-32, Table C.2), suggesting that both were deposited during Group I glaciation. The profile can be interpreted in two ways; (1) granite till in northern Eureka Sound may represent a deglacial phase following retreat of central

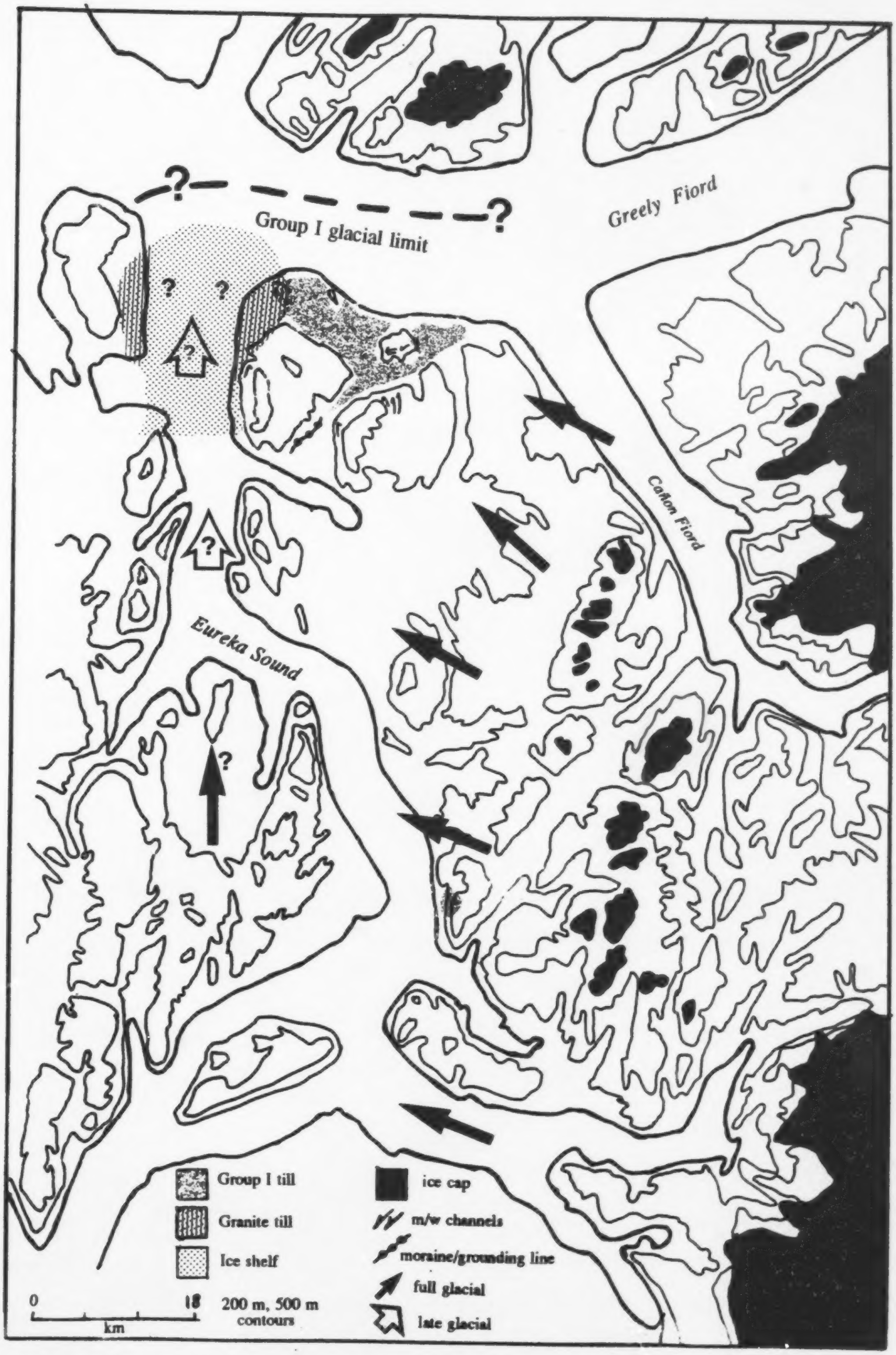

?
?
Group I glacial limit
Greely Fiord
? ?
?
?
Cañon Fiord
Eureka Sound
?
Group I till
Granite till
Ice shelf
ice cap
m/w channels
moraine/grounding line
full glacial
late glacial
0
18
km
200 m, 500 m contours

Figure 4.5: Proposed paleogeography of western Fosheim Peninsula duringGroup I glaciation. The proposed ice shelf in northern Eureka Sound represents a late glacial advance.

Ellesmere ice, or (2) during full glacial conditions, a stronger central Ellesmere ice flow coalesced with and diverted Eureka Sound ice around Troelsen ridge. This latter interpretation is appealing because it incorporates Eureka Sound ice into the ice-maximum configuration. Otherwise, Eureka Sound ice, originating on southeastern Ellesmere Island, must be advancing into Greely Fiord while central Ellesmere Island ice was retreating. Ironically, this interpretation (#1 above) is more consistent with the field evidence. The upper limit of granite erratics on Troelsen ridge is interpreted to reflect the surface profile of Eureka Sound ice as it entered Greely Fiord. The maximum thickness of ice in the northern part of the sound is estimated to be 725 m (275 m asl + *ca.* 450 m water depth; see Fig. 3.16). Given that Greely Fiord is more than 800 m deep off Iceberg Point and that deglacial sea levels were likely higher, then any late glacial advance of Eureka Sound ice must have floated on entering the fiord, producing an ice-shelf. A strong drawdown of ice is reflected in the granite profile on Troelsen ridge.

Subdued ridges composed of granite-rich till were mapped on the western margin of Iceberg Point lowlands (Figs. 3.19 and A.5). Given the location of this site in relation to the proposed former ice shelf in northern Eureka Sound, these landforms are tentatively interpreted as grounding lines that were formed during pinning of the ice shelf on the submerged lowlands.

The occurrence of fossiliferous, non-granite Group I till in central Eureka Sound (site 32) places the eastern margin of northward-flowing ice on Axel Heiberg Island. Either local ice from the Sawtooth Range or regional ice from farther east occupied central Eureka Sound at this time. The apparent absence of fossiliferous till elsewhere along this coast may have broader implications for the evolution of ancestral Eureka Sound (discussed later).

Ice retreat from Group I glaciation is recorded by high-elevation meltwater channels on northern Black Top Ridge (up to 600 m asl) and Snow Goose hill (Fig. 4.5). Together they indicate an eastward and southeastward retreat of ice across the peninsula. Meltwater channels on western Black Top Ridge (up to 280 m asl) and a prominent kame moraine on the adjacent lowlands record the southerly retreat of ice towards Slidre Fiord. It has already been proposed that, during deglaciation, Eureka Sound ice advanced into Greely Fiord and floated as an ice shelf. The thickness of this ice shelf is estimated to be 145 m using the upper limit of granite-rich till on Iceberg Point lowlands (275 m asl) as a maximum elevation on the ice-shelf surface, and the elevation of grounding lines on the lowlands (130 m asl) to represent the base of the ice shelf. Therefore, the maximum sea level required to float the ice shelf is 260 m asl. Although this is only a maximum sea level estimate, M. Marsden (pers. comm. 1990) reported 'old' raised beaches more than 200 m asl on the western slope of Troelsen ridge. These beaches have not been verified but their reported elevation is at least 50 m above the Holocene marine limit and therefore, it is unlikely that they were confused with 'younger' shoreline features. Other evidence of high sea levels on the peninsula cannot be exclusively related to Group I deglaciation.

4.2.3 Group II glaciation

The absence of shells with lower aIle/Ile ratios from northwestern Fosheim Peninsula suggests that this area lay beyond the limit of any subsequent ice advance. Based on the distribution of till samples with exclusively lower ratios or mixed ratios, the minimum extent of Group II glaciation is roughly located between Slidre Fiord and Cape Lockwood (Fig. 4.6). The only landform that possibly records this ice limit is a dissected delta at the northern tip of Black Top Ridge which was deposited into a former sea level at *ca.* 170 m asl, and prograded across Group I fossiliferous till (site 30). This delta is attributed to Group II glaciation, and not Group I deglaciation, based solely on its contemporary sea level. Elsewhere on the peninsula, sea levels associated with Group II glaciation are in the range of 170-184 m asl (Fig. 4.6).

Ice flow patterns appear similar to Group I glaciation, with Eureka Sound ice flowing north and coalescing with extensive local ice and/or regional ice from central Ellesmere Island. Ice thickness in Eureka Sound near Hare Cape ridge must have exceeded 1000 m in order to deposit shells and granite erratics on the local summit. However, farther north across Slidre Fiord, glacial deposits are non-fossiliferous and do not contain granite. This abrupt change is difficult to explain in terms of ice dynamics, flow patterns or topography and therefore, it is cautiously interpreted to represent the limit of Group II glaciation. Because this limit is poorly defined and most landforms cannot be dated, there is some uncertainty regarding the relative age of landforms on north-central Fosheim Peninsula (north of Slidre River).

Ice retreat from Group II glaciation is recorded by moraines, meltwater channels and outwash deposits (Fig. 4.6). The delta at 170 m asl, marking the glacial limit on the northern tip of Black Top Ridge, was incised by meltwater, causing reworking and regrading of outwash terraces to lower deglacial sea levels. Pre-existing, ice-marginal channels on Black Top Ridge were either re-occupied or cross-cut by younger channels as ice retreated from the basin; hence, the complex channel pattern on the ridge (Figs. A.1 and A.2).

Farther east, mesas and flat-topped terraces appear to form a dissected plain grading from 250 to 175 m asl (Fig. 4.6). These surfaces are thought to be younger than the Neogene high terraces although they may be composed of recycled high terrace sediment. Their abrupt lower limit, together with their terrace morphology, imply that they are remnants of a delta which graded to a former sea level at 175 m asl. During deglaciation, central Ellesmere ice stood at the escarpment on the central lowlands (Fig. A.3), producing meltwater which eroded proglacial channels on the uplands and fed the delta farther north (Fig. 4.6).

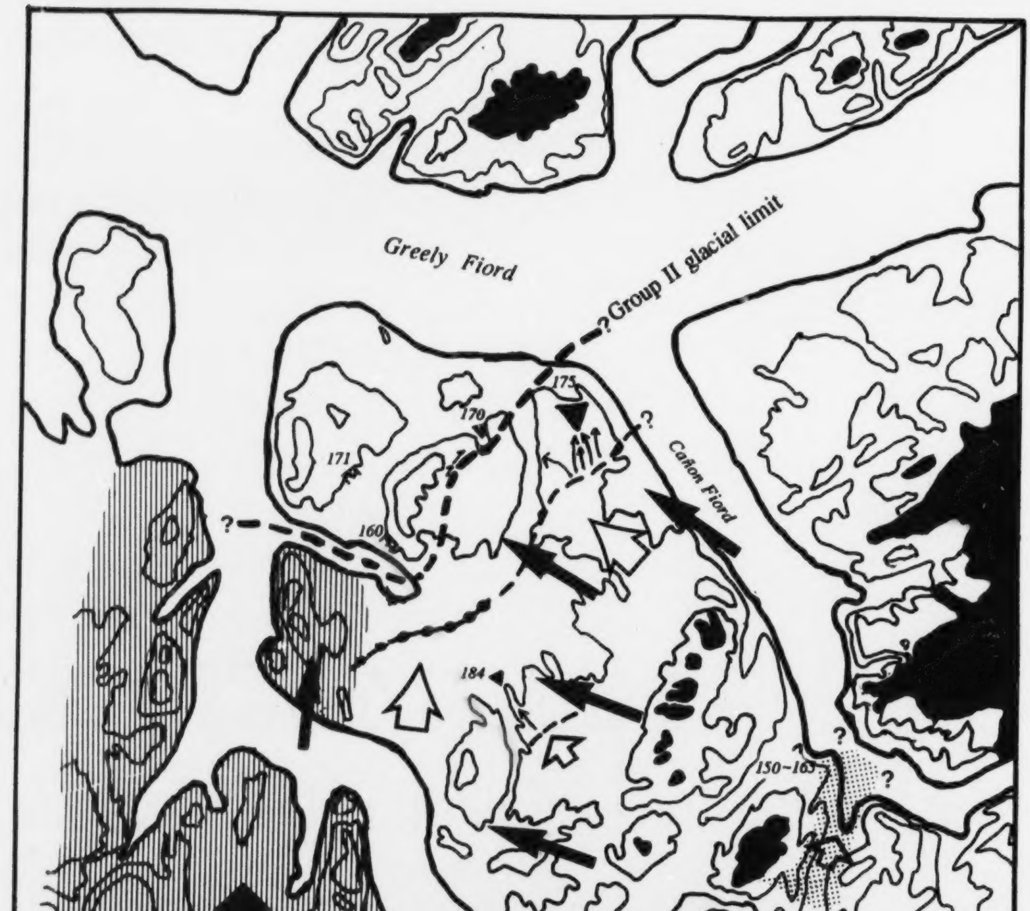

Greely Fiord
?Group II glacial limit
175
170
171
160
?
Cañon Fiord
184
150–165
?
?
?
Granite till
Ice shelf
ice cap
washing limit
delta
full glacial
deglacial
Moraine/grounding line
recessional stage
200 m, 500 m contours
0
18

Figure 4.6: Proposed paleogeography of western Fosheim Peninsula during Group II glaciation. Recessional ice margins relate to specific features and cannot be traced on the ground.

Romulus moraine (114-134 m asl), at the head of Slidre Fiord, marks ice recession during Group II deglaciation. Amino acid ratios on ice-transported shells both from inside the moraine and from adjacent Hare Cape ridge are similar, suggesting that the moraine was deposited during the same glacial episode (compare sites 33-36 and site 38, Table C.2). At one location on the lowlands, the surface till is underlain by interbedded tills and delta/fan deposits (Section 4, Appendix B). The delta/fan deposits indicate substantial crustal depression (90-100 m asl) during glacial buildup prior to the arrival of ice at the site. The lower tills may be assigned to either Group I or Group II glaciations; however, the surface till contains shells with exclusively lower amino acid ratios which can only be related to Group II glaciation. The form of Romulus moraine and the lithology of its surficial sediments indicate that it was deposited at the margin of coalescent ice from Eureka Sound, central Ellesmere Island and Fosheim Peninsula.

Outwash in upper Big Slide creek valley was deposited by meltwater from an ice margin retreating southward across the peninsula (Fig. 4.6). The outwash graded to a delta recording former sea level at 184 m asl (Fig. 4.6). Eureka Sound ice built a lateral moraine at 225 m asl on the southern end of the Sawtooth Range (Fig. 4.6). Although fossiliferous till, with exclusively higher amino acid ratios exists at the same elevation farther north (site 32), the moraine probably represents a recessional margin, especially considering the ice thickness (>1000 m) at Hare Cape ridge during the maximum of Group II glaciation. A till sample (site 40) downslope from the moraine contained a mixture of shell ages, demonstrating that the older fossiliferous till had been reworked at lower elevations. Sloan (1990) described a recessional stage of a regional glaciation in central Cañon Fiord where a trunk glacier from southeastern Fosheim Peninsula floated in a relative sea level at 150-165 m asl, that is, 30-45 m above the Holocene marine limit. Marine shells associated with this recessional stage dated >29 000 BP and were considered to be non-finite (Sloan 1990).

4.2.4 Regional correlation and absolute age estimates

To facilitate regional comparisons of amino acid data, it is important to select areas that have experienced similar thermal histories. Of most importance is the MAT and the extent of former ice cover which can have a significant effect on the rate of racemization. Of lesser importance, is the sea level history which, as suggested above, may not be as critical. Considering that the MAT at Eureka is -20°C, regional correlations are confined to northern Greenland (-16°C MAT), northeastern Ellesmere Island (-18°C MAT) and northwestern Ellesmere Island (≤-20°C). Only broad correlations with the Alaskan Arctic Coastal Plain (-10° to -13°C) are proposed. Meaningful correlations cannot be made with northwestern Greenland (Sejrup 1990; Funder and Houmark-Nielsen 1990) because the MAT is only -9°C nor with the central Canadian arctic where a continuous ice cover existed for at least all of Late and most of Middle Wisconsinan time (Dyke and Matthews 1987).

Ice-transported shells reflect the age of sediments incorporated during glacial advance; therefore, the lowest amino acid ratio represents the maximum relative age of glaciation. This ratio also provides a minimum relative age for the preceding marine episode. Commonly, however, only mean ratios are published for marine transgressions and therefore, this value is used to construct a regional framework. As ratios in the total acid hydrolysate are considered to have the best time-resolution in extreme environments (Miller and Brigham-Grette 1989), the following discussion focuses on these data.

Group I marine episode

With the exception of some shell fragments from site 27, group I samples have a limited range of amino acid ratios with a mean of 0.127 (n=17, CV=25%; Table 4.1). Members A and B of the Kap København Formation on northern Greenland are fossiliferous marine units that have combined mean ratios of 0.156 ± .033 (Funder *et al.* 1987) and 0.13 (Funder *et al.* 1985). Although ratios for individual members are difficult to interpret, it is proposed that Member A is the most likely equivalent of Group I shells (Table 4.2). Member A was deposited in a higher relative sea during the advance of the Greenland Ice Sheet whereas Group I shells were eroded and redeposited by an extensive regional glaciation on Ellesmere Island (see next section).

Member A of the Kap København Formation is dated between 2.4 and 3.2 Ma based on faunal, paleobotanical, paleomagnetic and amino acid ratios (Table 4.2). Brigham-Grette and Carter (1992) correlated this member with the Bigbendian transgression on the Alaskan Arctic Coastal Plain (Table 1.1). Its age is estimated to be *ca.* 2.48 Ma. The marine sediments at White Point probably represent an equivalent marine transgression on northwestern Ellesmere Island based on their faunal and foraminiferal content (2.4-3.2 Ma); however, two amino acid ratios suggest a younger age (0.054; Brigham-Grette and Carter 1992). In summary, it is proposed that group I shells from western Fosheim Peninsula represent a marine episode predating the onset of regional, and probably, northern hemisphere glaciation *ca.* 2.5 Ma (Table 4.2). Shells with anomalous ratios (0.3 total, site 27C) from group I are likely pre-Pleistocene.

Group I glaciation

The lowest amino acid ratios from group I shells in the study area (*free* 0.372; *total* 0.089) are roughly similar to ratios from the Greenland till on northeastern Ellesmere Island (*free* 0.43, *total* 0.125). However, as the latter ratios are based on two

Table 4.2: Tentative correlation of late Pliocene/early Pleistocene glaciations and marine episodes using total amino acid ratios (*italics*) and absolute age estimates. See text for discussion.

WESTERN FOSHEIM PENINSULA -20°C MAT This study	**NORTHWESTERN ELLESMERE ISLAND** Brigham-Grette and Carter 1992	**NORTHEASTERN ELLESMERE ISLAND** -18°C MAT Retelle 1986, Lemmen and England 1992	**NORTHERN GREENLAND** -16°C MAT Funder *et al.* 1985, 1987	**ALASKAN ARCTIC COASTAL PLAIN** -10° to -13°C MAT Brigham-Grette and Carter 1992
0.127 Group I marine episode	*0.054* WhitePoint 2.4 - 3.2 Ma	?	*0.13, 0.156* Kap København Member A 2.4 - 3.2 Ma	*0.126, 0.15* Bigbendian 2.48 Ma
0.089 Group I glaciation		0.125 GREENLAND 0.4- 1 Ma	ICE ADVANCE 2 - 2.4 Ma	
0.065 Group II marine episode		*0.063* Robeson aminozone >70 Ka	Kap København Member B 2 Ma	*0.085, 0.09* Fishcreekian 2.14 - 2.48 Ma
0.047 Group II glaciation				

unidentified fragments, a direct correlation must be tentative. England (1987) suggested that the Greenland ice advance onto Ellesmere Island could be as old as 1 Ma, based on broad comparisons with amino acid ratios from Alaska, whereas Lemmen and England (1992) proposed a provisional age estimate of 0.4-0.6 Ma. Here, it is suggested that these dates are underestimated and that the Greenland ice advance and Group I glaciation are coeval with the expansion of the Greenland Ice Sheet on northern Greenland (Table 4.2). This correlation is based on the similarity in amino acid ratios from sediments associated with the three events. Member A of the Kap København Formation has the most reliable age estimate between 2.4 and 3.2 Ma although the timing of the glacial maximum may have coincided with the first Late Cenozoic northern Hemisphere glaciation at *ca.* 2.5 Ma. The regional synchroneity and age of deglacial marine sediments support this correlation (next section).

Group II marine episode

When Greenland ice retreated off northeastern Ellesmere Island, local ice responded by advancing as ice shelves into Nares Strait (J. England, pers. comm. 1992). Shells related to ice shelf occupation provided mean aIle/Ile ratios of 0.05. The Robeson aminozone includes all samples related to the marine transgression during retreat of Greenland ice from Ellesmere Island and has a mean ratio of 0.063 (Retelle 1986). Group II shell samples from western Fosheim Peninsula have a discrete range of amino acid ratios with a mean of 0.065 (n=20, CV=17%). Based on the similarity of amino acid ratios, it is concluded that Group II marine episode represents a deglacial high sea level that was regionally synchronous following Group I glaciation (Table 4.2).

The elevation of this transgression on northeastern Ellesmere Island was measured at 285 m asl (England *et al.* 1981). This is similar to the limit of the transgression represented by Member B of the Kap København Formation on northern Greenland (230 m asl; Bennike and Böcher 1990). Considering that both shorelines immediately postdate an extensive advance of Greenland ice, it is not unreasonable to assume that they represent the same marine transgression. Subsequent transgressions on northeastern Ellesmere Island did not reach such an elevation (early Holocene = *ca.* 120 m asl; England 1978). As the age of Member B is estimated to be 2 Ma (Funder *et al.* 1985), this implies that both the Robeson aminozone and Group II marine episode have a similar age (Table 4.2). This regional transgression may also be correlated with the Fishcreekian transgression on the Alaskan Arctic Coastal Plain. Fishcreekian sediments contain Canadian Shield erratics which probably signify the breakup of the earliest regional ice sheet and the subsequent ice-rafting of glacial debris. The age of this transgression is estimated to be 2.14-2.48 Ma (Brigham-Grette and Carter 1992).

Group II glaciation

The lowest total ratio from Group II till samples (0.047) is similar to the lowest mean ratios recorded for the Robeson aminozone (0.052; Retelle 1986). This suggests that either Group II glaciation immediately followed Group I deglaciation or group II shell samples include only raised marine sediments that relate to Group I deglaciation. The former interpretation implies that Group II glaciation is only slightly younger than 2 Ma whereas the latter interpretation can only suggest a pre-last glacial age. Till samples with similar ratios have not been observed in other areas; however, Lemmen and England (1992) tentatively recognized a post-Greenland ice advance on northeastern Ellesmere Island which they suggested was probably pre-Wisconsinan. The lowest amino acid ratio associated with this event is 0.02.

4.2.5 Discussion

During the last glaciation of western Fosheim Peninsula, ice caps grew on ice-free uplands and cirque glaciers advanced several kilometres. In contrast, older glaciations were much more extensive (*research objective #1*). Meltwater channels, erratics distribution and fossiliferous diamicton demonstrate that ice once filled Eureka Sound and overtopped the highest summits of the peninsula. The most extensive glaciation, Group I, inundated the peninsula and coincided with the expansion of the Greenland Ice Sheet onto northeastern Ellesmere Island. It is suggested to be *ca.* 2.5 Ma. The subsequent Group II glaciation was also extensive but the northwestern part of the peninsula remained ice-free. The age of Group II glaciation is uncertain but it may be as old as 2 Ma. Such extensive glaciations indicate a significant departure from last glacial conditions.

High arctic glacial climate

Climate during most of the last glaciation was colder and drier than present (Bradley and England 1978); however, late ice buildup is attributed to a marked warming and an increase in precipitation after 14 ka BP (England 1992) which may have been accentuated by a reduction in the sea ice cover, providing increased precipitation. This local climatic effect maintained upland ice caps on the peninsula through the mid-Holocene. Because precipitation is regarded as the critical climate control on modern glacier growth in the high arctic, it has been argued that regional moisture availability also dictated the extent of older glaciations (Bradley and England 1978; Miller and de Vernal 1992).

Raymo *et al.* (1990) have shown, through computer modelling, that the Arctic Ocean sea ice cover greatly diminishes moisture availability in the high arctic and that a reduced sea ice cover would produce significant increases in winter precipitation.

Therefore, it is proposed that extensive glaciations of northern and western Ellesmere Island must have coincided with a greatly reduced Arctic Ocean sea ice cover (*cf.* Lemmen and England 1992). To test this relationship, a generalized chronology of high arctic glaciation is compared to the Arctic Ocean sea ice history.

Herman and Hopkin's (1980) interpretation of Arctic Ocean history is used here because it is the one that appears to be in closest agreement with the terrestrial geologic record (see section 1.4.2. for discussion of alternative records). Group I glaciation, and the maximum advance of Greenland ice onto Ellesmere Island, are tentatively assigned an age of *ca.* 2.5 Ma. Raised marine records on northern Greenland (Member A, Kap København Formation) and the Alaskan Arctic Coastal Plain (Bigbendian transgressive sediments) demonstrate that there were at least seasonal ice-free conditions along the margin of the Arctic Ocean at this time (Funder *et al.* 1985; Carter *et al.* 1986). Herman and Hopkins (1980, p. 209) argued that at 2.5 Ma, Arctic Ocean water temperature was "incompatible with the presence of a perennial Arctic ice pack, but a seasonal ice cover may have been present." Group I deglaciation and retreat of the Greenland Ice Sheet coincides with the maximum of the early Pleistocene warm period at 2 Ma (*cf.* Repenning 1990). Associated marine sediments provide evidence for a warm Arctic Ocean without perennial sea ice cover (Fishcreekian marine sediments) and a forest tundra vegetation in the high arctic (Member B, Kap København Formation). The correlation between extensive glacier growth and at least a seasonally ice-free Arctic Ocean appears to be strong for Group I glaciation; however, the breakup of this ice cover during a period warmer than present demonstrates the sensitivity of this environment to climate change and reinforces the climatic significance of an ice-free Arctic Ocean (*cf.* Ewing and Donn 1956).

The timing of Group II glaciation is uncertain but, according to the proposed hypothesis, it should predate a perennial Arctic Ocean sea ice cover which originated between 0.4 and 0.7 Ma (Clark 1990; Herman and Hopkins 1980). The increased circulation between the Arctic Ocean and the North Greenland Sea, which likely caused the early Pleistocene warm period, is still evident in the ocean sediment record at 1 Ma (Clark 1990). Biological productivity suggests that this period may have experienced warmer ocean conditions than had existed prior to, or since this time (Clark 1982; Gilbert and Clark 1983). It is suggested, therefore, that climate and sea ice conditions were most favourable for Group II glaciation immediately prior to the formation of a perennial Arctic Ocean sea ice cover at 0.4-0.7 Ma.

This chronology suggests that the last glaciation (Late Wisconsinan / early Holocene) in the high arctic was more extensive than any other Late Pleistocene glaciation. Although the extent of the last glaciation on western Fosheim Peninsula is attributed to reduced sea ice cover in *local* channels and fiords, it reaffirms the importance of precipitation for glacier growth in the high arctic (*cf.* Bradley and England 1978).

Late Tertiary landscape

The glacial and sea level history of western Fosheim Peninsula provides information on the late Tertiary regional landscape (*research objective #4*). Two different landscape models have been proposed for west-central Ellesmere Island. England (1987) argued that if there was ever a regional ice sheet over the high arctic, that it occurred on a prefaulted, contiguous landmass during the late Tertiary. The absence of glacial geologic evidence beyond the last ice limit in Greely Fiord prompted England (1987) to speculate on the tectonic origin of high arctic fiords which eventually terminated the possibility for regional glaciations comparable to the earliest events. In contrast, Ricketts (1987*b*) proposed that the present valleys and fiords evolved from an initial consequent drainage that paralleled the late Tertiary geologic structure (that is, parallel to the ridges and valleys of eastern Fosheim Peninsula) and which was later overdeepened by Pleistocene glaciation. Where fiords and interisland channels cross-cut the structural grain (*e.g.* Cañon Fiord), Ricketts (1987*b*) points to structural discordances as the controlling factor in drainage evolution.

The identification of paleodrainage systems on eastern Fosheim Peninsula (Fig. 2.11) suggests that the present drainage may be regarded as a subsequent system which developed in response to the formation of fiords and interisland channels. During early Paleocene to middle Eocene, consequent rivers carried fine-grained sediment from the Precambrian Shield uplands (southeastern Ellesmere Island; Figs. 1.3 and 2.5) to the Fosheim Basin, where it was deposited as the Eureka Sound Group. When the Eurekan Orogeny reached its climactic phase, rivers continued to flow to the Fosheim Basin as an antecedent drainage through the uplifting Sawtooth Range. These river valleys are preserved as wind gaps in the modern Sawtooth Range (Fig. 2.11). During this period, broad peneplains were likely formed on eastern Axel Heiberg Island and high terrace sediments were deposited as an extensive, northwestward sloping fluvial plain on western Fosheim Peninsula (Fig. 2.9). These Neogene sediments demonstrate that rivers continued to flow through the Sawtooth Range until such a time when newly formed fiords and marine channels captured and truncated the antecedent drainage. In response to regional uplift, peneplains, fluvial plains and the new subsequent drainage basins were deeply incised, isolating Neogene fluvial sediments more than 300 m above present valleys (Figs. 2.13 and 4.6). It is not surprising therefore, that the youngest organic remains (early Quaternary?) have been recovered from high terrace sediments in a subsequent drainage system (Wolf Valley, Fig. 2.11; *cf.* Matthews and Ovenden 1990). According to this chronology, some fiords and marine channels must be coeval with the development of subsequent drainage and the capture of antecedent drainage. Hence, the recognition and dating of remnant fluvial deposits associated with these discrete drainage systems should provide critical age estimates on the formation of local fiords and interisland channels.

The configuration of older, extensive glaciations and the proposed chronology of glacial and marine episodes on western Fosheim Peninsula provide some control on the landscape evolution of this region. Group I shells demonstrate that Greely Fiord and northern and south-central Eureka Sound existed as marine embayments *prior* to the extensive Group I glaciation at *ca.* 2.5 Ma. If this was the first extensive glaciation of west-central Ellesmere Island, as is assumed in this chronology, then it must postdate the breakup of the contiguous Fosheim Basin. Consequently, Greely Fiord and Eureka Sound cannot have a glacial origin although they may have been modified by subsequent glaciations. Lemmen (1990) also concluded this for Disraeli Fiord, northern Ellesmere Island. The preservation of *in situ*(?) marine sediments at White Point that are tentatively correlated with Group I marine episode, suggests that Nansen Sound may never have been completely inundated by ice.

The 'high terraces' on western Fosheim Peninsula are composed of the youngest sediments associated with the formerly contiguous Fosheim Basin. Organic beds in these terraces represent a coniferous forest which, on the basis of diagnostic plant fossils, may be equivalent to the Meighen Island Beaufort Formation, dated 3 Ma (J. Matthews, pers. comm. 1991). Considering that Group I glaciation is estimated to be *ca.* 2.5 Ma, this provides a relatively short period for Eureka Sound to evolve solely by tectonic processes. To what degree Eureka Sound has been modified by glacial erosion is unknown.

CHAPTER FIVE

Summary

Western Fosheim Peninsula is located near the centre of the Eureka Sound lowlands on west-central Ellesmere Island. It is isolated from the Arctic Ocean and Baffin Bay by mountain ranges which protect it from the incursion of cold air masses and cyclonic activity, respectively. As a result, it experiences anomalously high summer temperatures for its latitude and has the lowest mean annual precipitation in Canada. This severe aridity is responsible for the limited ice cover on the peninsula which is confined to small cirque glaciers in the Sawtooth Range. The local bedrock is dominated by poorly consolidated, sedimentary rocks of the Sverdrup Basin.

The surface materials of western Fosheim Peninsula are divided into six genetic units. Weathered bedrock/Residuum forms the largest unit. Although the local bedrock is poorly indurated and strongly susceptible to weathering, tors and other features suggest a substantial period of subaerial exposure. Glacial erratics are widespread but glacial landforms are preserved only on more resistant bedrock. Ice-marginal meltwater channels are common, recording the systematic retreat of former ice cover. Remnants of Neogene high terraces are preserved on the peninsula, armoured by a gravel lag. This lag is interpreted, in some cases, to be a winnowed till but in others, it may be of glaciofluvial origin. High terrace sediments are treated as a surficial unit because they form conspicuous mesas and buttes. Till, mainly as a veneer, is the second largest surficial unit. Till distribution is described according to erratic lithology. Granite-rich till which forms a long, narrow dispersal train roughly coincident with Eureka Sound, probably originated in the Canadian Shield of southeastern Ellesmere Island. The remaining till was deposited by either regional ice from central Ellesmere Island or local ice caps and cirque glaciers. Meltwater channels and moraines document former ice margins and deglacial patterns. Glaciofluvial sediments cover a relatively small area. They form kame moraines and outwash fans and terraces, representing ice-marginal and proglacial environments, respectively. Raised marine sediments of deltaic, nearshore and beach facies comprise the third most extensive surficial unit. Small deltas record Holocene marine limit whereas thick glaciomarine sediments document the history of glacial runoff. Postglacial marine sediments, including small emerged deltas, thin nearshore silt and, in places, extensive beach gravel, record Holocene emergence. There is limited evidence for pre-Holocene sea levels. Fluvial sediments are divided into active and inactive alluvium; both commonly occur together though neither is extensive.

Discontinuous moraines and meltwater channels that either encircle broad uplands or occupy the lower valleys of cirque basins outline a restricted ice cover during the last glacial maximum. In all cases, these landforms clearly cross-cut weathered bedrock and older glacial deposits. The paleoglaciation level at this time is estimated to have been 700-800 m lower than present. Although the severe aridity of northern and central Ellesmere Island was a primary climatic constraint on the magnitude of the last glaciation (Bradley and England 1978), increased local moisture availability led to a more extensive local ice cover than that predicted from present climate characteristics.

During the last glacial maximum, some glaciers reached the sea which occupied broad lowlands on the peninsula. Regional isobases drawn on the limit of this 'full glacial' Innuitian Sea (*cf.* England 1990) outline a ridge of emergence centred over an area that experienced the smallest ice load change during the last glaciation. This pattern is unlike those from other parts of Arctic Canada and may reflect rheological, glacioisostatic, or neotectonic factors that are not fully understood. The age of this shoreline Has not been determined for the study area; however, England (1992) proposed that a stable sea level occupied marine limit in inner Greely and Cañon fiords between 7800 and 8600 BP.

During local deglaciation, sediment-laden meltwater reached the sea, depositing thick sequences of glaciomarine sediments in submerged river valleys. High sedimentation rates were sustained through the mid-Holocene, demonstrating that upland ice caps experienced slow deglaciation during the warmest period of the Holocene (*cf.* Bradley 1990). This apparent contradiction emphasizes the importance of precipitation for the growth of high arctic glaciers (*cf.* Bradley and England 1978). The increased warming of the early Holocene led to reduced sea ice conditions which, in turn, produced precipitation that maintained local ice caps. Similar conditions were reported for Borup Fiord (England 1990, 1992).

The postglacial emergence history of western Fosheim Peninsula has not been resolved by this study and will be pursued in future research. A maximum reconstruction of Holocene emergence has been proposed by England (1992) for adjacent Greely Fiord; however, the apparent absence of shells close to marine limit on western Fosheim Peninsula may reflect local paleoenvironmental conditions that constrained the entry of marine organisms during deglaciation.

Beyond the last ice limit in the study area, there is abundant evidence of older, more extensive glaciations. The occurrence of two tills with ice-transported shells of discrete relative ages, as determined by amino acid ratios, can only be reasonably explained in terms of two older glaciations. These are provisionally named Group I and Group II glaciations for the oldest and youngest events, respectively.

Group I glaciation inundated Fosheim Peninsula and was likely responsible for the deposition of erratics on the highest summits, adjacent to Greely Fiord. The limit of this advance is unknown but the distribution of erratics suggests that it was at least 800 m thick on the northern peninsula. During either full glacial or late glacial conditions, ice from central and southeastern Ellesmere Island coalesced in Eureka Sound. The upper limit of granite-rich till in northern Eureka Sound records a steep ice profile which may indicate that Eureka Sound ice floated upon entering Greely Fiord during a late glacial advance. Meltwater channels and a kame moraine record an eastward and southeastward retreat of ice across the northern peninsula.

Fossiliferous till associated with this ice advance contains shells that have similar amino acid ratios to the oldest marine sediments of the Kap København Formation, northern Greenland, dated 2.4-3.2 Ma. Member A of this formation was deposited during the advance of the Greenland Ice Sheet. These ratios are also similar to shell fragments from granite-rich till deposited by the maximum advance of the Greenland Ice Sheet onto northeastern Ellesmere Island. Based solely on the correlation of amino acid ratios, it is suggested that Group I glaciation was coeval with the maximum advance of the northwestern Greenland Ice Sheet and that this advance redeposited shells related to the preceding Group I marine episode, tentatively dated 2.4-3.2 Ma. Therefore, the maximum age of Group I glaciation is possibly 2.4 Ma which may coincide with the first Late Cenozoic northern hemisphere glaciation.

Retreat of Greenland ice from northeastern Ellesmere Island and the subsequent advance of Ellesmere ice into Nares Strait were associated with shells that have similar amino acid ratios to Group II marine episode on western Fosheim Peninsula. It is proposed that deglacial shorelines on northeastern Ellesmere Island may be equivalent to Member B marine transgression of the Kap København Formation, based solely on elevation (285 m *vs.* 230 m). Therefore, these correlations provide an age estimate of *ca.* 2 Ma for Group I deglaciation.

The minimum extent of Group II glaciation is outlined by the distribution of group II shell samples which consist exclusively of lower amino acid ratios. It is tentatively positioned between Slidre Fiord and Cape Lockwood. A delta at 170 m asl near northern Black Top Ridge may represent the limit of this advance. During Group II glacial maximum, ice from Eureka Sound, central Ellesmere and Fosheim peninsula coalesced along the western fringe of the study area. Moraines, meltwater channels and outwash deposits record the retreat of ice across the peninsula in conjunction with higher relative sea levels between 170 and 185 m asl.

The maximum age of Group II glaciation is estimated to be *ca.* 2 Ma; however, it may be as young as 0.4-0.7 Ma based on paleoclimatic considerations. Lemmen and England (1992) recognized a pre-Wisconsinan glaciation on northeastern Ellesmere Island that also postdated the maximum advance of the Greenland Ice Sheet; however, according to their chronology, it could be younger than 0.4 Ma.

Older, extensive glaciations of western Fosheim Peninsula represent a significant departure from last glacial or present environmental conditions. Because precipitation is regarded as the critical climate control on modern glacier growth, it has been argued that more extensive glaciations must have been favoured by much greater precipitation (Bradley and England 1978; Miller and de Vernal 1992). The proposed relationship between regional glaciation of the high arctic and a greatly reduced Arctic Ocean sea ice cover (*cf.* Lemmen and England 1992) appears to be consistent with the geological evidence for Group I glaciation. However, the tentative correlations on which this test is based need to be independently verified and a more rigorous chronology developed.

The glacial history of the study area is used to constrain landscape evolution models for west-central Ellesmere Island (England 1987; Ricketts 1987). Group I shells demonstrate that Greely Fiord and parts of Eureka Sound existed as marine embayments prior to Group I glaciation; therefore, regional glaciation must postdate the breakup of a contiguous Fosheim Basin (*i.e.* at least nascent fiords and marine channels predate extensive glaciation of this area). It is proposed that the detailed mapping and dating of Neogene fluvial sediments, which record the late Tertiary drainage evolution of the peninsula, may provide proxy data on the age and evolution of the surrounding fiords and marine channels.

REFERENCES

Alt, B. T. 1979. Investigations of summer synoptic controls on the mass balance of the Meighen Ice Cap. Atmoshpere-Ocean, **3:** 181-199.

Alt, B. T., and Maxwell, J. B. 1990. The Queen Elizabeth Islands: A case study for arctic climate data availability and regional climate analysis. In: *Symposium on the Canadian Arctic Islands, Canada's missing dimension*, Harrington, C. R., ed., National Museum of Natural Sciences, Ottawa.

Andrews, J. T. 1970. A geomorphic study of postglacial uplift with particular reference to Arctic Canada. Institute of Briish Geographers, Special Publication 2, London, 156p.

Andrews, J. T., and Miller, G. H. 1972. Quaternay history of northern Cumberland Peninsula, Baffin Island, N.W.T., Part IV: maps of the present glaciation limits and lowest equilibrium line altitude for north and south Baffin Island. Arctic and Alpine Research, **4:** 45-59.

Asudeh, I., Forsyth, R., Stephenson, R., Embry, A., Jackson, H. R., and White, D. 1989. Crustal structure of the Canadian polar margin: results of the 1985 seismic refraction survey. Canadian Journal of Earth Sciences, **26:** 853-866.

Atmospheric Environment Service 1982. Canadian climate normals 1951-80. Temperature and precipitation: The North - Yukon Territoty and Northwest Territories. Environment Canada, Downsview, Ontario.

Axelrod, D. I. 1984. An interpretation of Cretaceous and Teriary biota in polar regions. Palaeogeography, Palaeoclimatology, Palaeoecology, **45:** 105-147.

Balkwill, H. R. 1978. Evolution of Sverdrup Basin, Arctic Canada. Bulletin of the American Association of Petroleum Geologists, **62:** 1004-1028.

Banham, P. H. 1977. Glaciotectonics in till stratigraphy. Boreas, **6:** 101-105.

Barron, E. J. 1985. Explanations of the Tertiary global cooling trend. Palaeogeography, Palaeoclimatology, Palaeoecology, **50:** 45-61.

Barry, R. G. 1989. The present climate of the Arctic Ocean and possible past and future states. In: *The arctic seas*, Herman, Y., ed., Van Nostrand Reinhold; New York, 1-46.

Basinger, J. F. 1986. Our Tropical Arctic. Canadian Geographic, **106:** 28-37.

Basinger, J. F. 1991. The fossil forests of the Buchanan Lake Formation (Early Teritary), Axel Heiberg Island, Canadian Arctic Archipelago: Preliminary floristics and paleoclimate. Geological Survey of Canada, Bulletin **403**: 39-65.

Bednarski, J. 1986. Late Quaternary glacial and sea level events, Clements Markham Inlet, northern Ellesmere Island, Arctic Canada. Canadian Journal of Earth Sciences, **23**: 1343-1355.

Bednarski, J. 1988. The geomorphology of glaciomarine sediments in a high Arctic Fiord. Géographie Physique et Quaternaire, **42**: 65-74.

Bednarski, J. 1990. An early Holocene Bowhead whale in Nansen Sound, Canadian Arctic Archipelago. Arctic, **43**: 50-54.

Bennike, O., and Bõcher, J. 1990. Forest tundra neighbouring the North Pole: plant and insect remains from the Plio-Pleistocene Kap København Formation, North Greenland. Arctic, **43**: 331-338.

Berger, A. L. 1984. Accuracy and frequency stability of the earth's orbital elements during the Quaternary. In: *Milankovitch and climate*, Berger, A. L.; Imbrie, J.; Hays, J.; Kukla, G.; Saltzman, B., eds., Reidel; Boston, 3-39.

Berger, W. H. 1982. Climate steps in ocean history - lessons from the Pleistocene. In: *Climate in Earth History*, Geophysics Study Committee (eds.), National Academy Press; Washington, D.C., 43-54.

Berggren, W. A. 1982. Role of ocean gateways in climate change. In: *Climate in Earth history*, Geophysics Study Committee (eds.), National Academy Press; Washington, D.C., 118-125.

Bergsma, B. M., Svoboda, J., and Freedman, B. S. 1984. Entombed plant communities released by a retreating glacier at central Ellesmere Island, Canada. Arctic, **37**: 49-52.

Bird, J. B. 1967. The Physiography of Arctic Canada (with special reference to the area south of Parry Channel). The John Hopkins Press; Baltimore, 336p.

Blake, W., Jr. 1970. Studies of glacial history in Arctic Canada. I. Pumice, radiocarbon dates, and differential uplift in the eastern Queen Elizabeth Islands. Canadian Journal of Earth Sciences, **7**: 634-664.

Blake, W., Jr. 1972. Climatic implications of radiocarbon-dated driftwood in the Queen Elizabeth Islands, arctic Canada. In: *Climatic changes in arctic areas during the last ten-thousand years*, Vasari, Y.; Hyvarinen, H.; Hicks, S., eds., University of Oulu, Oulu, Finland, 77-104.

Blake, W., Jr. 1975. Radiocarbon age determinations and postglacial emergence at Cape Storm, southern Ellesmere Island, Arctic Canada. Geografiska Annaler, **57A**: 1-71.

Blake, W., Jr. 1981. Neoglacial fluctuations of glaciers, southeastern Ellesmere Island, Canadian Arctic Archipelago. Geografiska Annaler, **63A**: 201-218.

Bliss, L. C., and Bliss, D. I. 1984. Polar deserts, their plant cover and plant production in the Canadian High Arctic. Holarctic Ecology, **7**: 305-324.

Boesch, H. 1963. Notes on the geomorphological history. In: *Preliminary Report 1961-1962*, Axel Heiberg Research Reports, McGill University; Montreal, 163-168.

Bornhold, B. D., Finlayson, N. M., and Monahan, D. 1976. Submerged drainage patterns in Barrow Strait, Canadian Arctic. Canadian Journal of Earth Sciences, **13**: 305-311.

Boulton, G. S. 1979. Processes of glacier erosion on different substrata. Journal of Glaciology, **23**: 15-38.

Bovis, M. J., and Barry, R. G. 1974. A climatological analysis of north polar desert areas. In: *Polar deserts and modern man*, Smiley, T. L.; Zumberge, J. H., eds., University of Arizona Press; Tucson, 23-31.

Boyer, S. J., and Pheasant, D. R. 1974. Delimiting of weathering zones in the fiord area of Eastern Baffin Island, Canada. Geological Society of America Bulletin, **85**: 805-810.

Bradley, R. S. 1975. ELA's, mass balance, and July freezing level heights in the Canadian High Arctic. Journal of Glaciology, **14**: 267-274.

Bradley, R. S. 1985. Quaternary Plaeoclimatology. Allen and Unwin; Boston, 472p.

Bradley, R. S. 1990. Holocene paleoclimatology of the Queen Elizabeth Islands, Canadian High Arctic. Quaternary Science Reviews, **9**: 365-384.

Bradley, R. S., and Eischeid, J. K. 1985. Aspects of the precipitation climatology of the Canadian High Arctic. In: *Glacial Geologic and Glacio-climatic Studies in the Canadian High Arctic*, Bradley, R. S., ed., Department of Geology and Geography, University of Massachusetts; Amherst, 250-271.

Bradley, R. S., and England, J. 1978. Recent climatic fluctuations of the Canadian High Arctic and their significance for glaciology. Arctic and Alpine Research, **10**: 715-731.

Bradley, W. C., Hutton, J. T., and Twidale, C. R. 1978. Role of salts in development of granitic tafoni, south Australia. Journal of Geology, **86**: 647-654.

Brigham-Grette, J., Matthews, J. V., Jr., and Marincovich, L., Jr. 1987. Age and paleoenvironmental significance of Arctica in the Neogene Beaufort Formation on Meighen Island, Queen Elizabeth Islands, Canada. In: *16th Arctic Workshop*, University of Alberta: Boreal Institute for Northern Studies; Edmonton, Alberta, 12-14.

Brigham-Grette, J., and Carter, L. D. 1992. Pliocene marine transgressions of Northern Alaska: Circumarctic correlations and paleoclimatic interpretations. Arctic, **45**: 74-89.

Brigham-Grette, J., and Kaufman, D. S. 1990. Latitudinal transect and aminostratigraphic correlation of Pliocene/Pleistocene marine beds, northern and western Alaska. Geological Society of America, Abstracts with Programs, **22**: A147.

Broecker, W. S., Kulp, J. L., and Tucek, C. S. 1956. Lamont natural radiocarbon measurements III. Science, **124**: 154-165.

Broecker, W. S., Peteet, D. M., and Rind, D. 1985. Does the ocean-atmosphere system have more than one stable mode of operation? Nature, **315**: 21-26.

Carter, L. D., Brigham-Grette, J., Marincovich, L., Pease, V. L., and Hillhouse, J. W. 1986. Late Cenozoic Arctic sea ice and terrestrial paleoclimate. Geology, **14**: 675-678.

Christie, R. L. 1967. Reconnaissance of the surfical geology of northeastern Ellesmere Island, Arctic Archipelago. Geological Survey of Canada, Bulletin 138.

Clark, D. L. 1982. Origin, nature, and world climate effect of Arctic Ocean ice cover. Nature, **300**: 321-325.

Clark, D. L. 1990. Arctic Ocean ice cover; geological history and climatic significance. In: *The Arctic Ocean region*, Grantz, A.; Johnson, L.; Sweeney, J. F., eds., Geological Society of America; Boulder, Colorado, 53-62.

Clark, D. L., Chern, L. A., Hogler, J. A., Mennicke, C. M., and Atkins, E. D. 1990. Late Neogene climate evolution of the central Arctic Ocean. Marine Geology, **93**: 69-94.

Clark, J. A., Farrell, W. E., and Peltier, W. R. 1978. Global changes in postglacial sea level: A numerical calculation. Quaternary Research, **9**: 265-287.

CLIMAP Project Members 1976. The surface of the ice age earth. Science, **191**: 1131-1144.

COHMAP 1988. Climatic changes of the last 18 000 years: observations and model simulations. Science, **241**: 1043-1052.

Cooke, R. U. 1979. Laboratory simulation of salt weathering processes in arid environments. Earth Surface Processes and Landforms, **4**: 347-359.

Cowan, E. A., and Powell, R. D. 1990. Suspended sediment transport and deposition of cyclically interlaminated sediment in a temperate glacial fjord, Alaska, U.S.A. In: *Glacial marine environments; Processes and sediments*, Dowdeswell, J. A.; Scourse, J. D., eds., Geological Society of London Special Publication 53, 75-90.

Craig, B. G., and Fyles, J. G. 1960. Pleistocene geology of Arctic Canada. Geological Survey of Canada, Paper **60-10**: 21 p.

Creber, G. T., and Chaloner, W. G. 1985. Tree growth in the Mesozoic and Early Tertiary and the reconsruction of paleoclimates. Palaeogeography, Palaeoclimatology, Palaeoecology, **52**: 35-60.

Davis, R. A., Finley, K. J., and Richardson, W. R. 1980. The present status and future management of Arctic marine mammals in Canada. Unpublished report, Northwest Territories Science Advisory Board, Yellowknife, 93p.

Dawson, M. R., West, R. M., Langston, W., Jr., and Hutchinson, J. H. 1976. Paleogene terrestrial vertebrates: northernmost occcurrence, Ellesmere Island, Canada. Science, **192**: 781-782.

Denton, G. H., and Hughes, T. J. 1981. The Last Great Ice Sheets. John Wiley & Sons; London, 484p.

De Paor, D. G., Bradley, D. C., Eisenstadt, G. E., and Phillips, S. M. 1989. The Arctic Eurekan orogen: A most unusual fold-and-thrust belt. Geological Society of America Bulletin, **101:** 952-967.

Devaney, J. R. 1991. Clastic sedimentology of the Beaufort Formation, Prince Patrick Island, Canadian Arctic Islands: late Tertiary sandy braided river deposits with woody detritus beds. Arctic, **44:** 206-216.

Domack, E. W. 1984. Rhymically bedded glaciomarine sediments on Whidbey Island, Washington. Journal of Sedimentary Petrology, **54:** 589-602.

Dowsett, H. J., and Cronin, T. M. 1990. High eustatic sea level during the middle Pliocene: evidence from the southeastern U.S. Atlantic Coastal Plain. Geology, **18:** 435-438.

Dunbar, C. O., and Rodgers, J. 1957. Principles of Stratigraphy. John Wiley; New York, 356.

Dyck, W., and Fyles, J. G. 1962. Geological Survey of Canada radiocarbon dates I. Radiocarbon, **4:** 13-26.

Dyck, W., and Fyles, J. G. 1963. Geological Survey of Canada Radiocarbon Dates II. Radiocarbon, **5:** 39-55.

Dyck, W., and Fyles, J. G. 1964. Geological Survey of Canada radiocarbon dates III. Radiocarbon, **6:** 167-181.

Dyck, W., Fyles, J. G., and Blake, W., Jr. 1965. Geological Survey of Canada radiocarbon dates IV. Radiocarbon, **7:** 24-46.

Dyke, A. S. 1979. Glacial and sea level history of southwestern Cumberland Peninsula, Baffin Island, N.W.T. Arctic and Alpine Research, **11:** 179-202.

Dyke, A. S. 1983. Quaternary geology of Somerset Island, District of Franklin. Geological Survey of Canada, Memoir 404.

Dyke, A. S., and Matthews, J. V., Jr. 1987. Stratigraphy and paleoecology of Quaternary sediments along Pasley River, Boothia Peninsula, Central Canadian Arctic. Géographie Physique et Quaternaire, **41:** 323-344.

Dyke, A. S., and Morris, T. F. 1988. Drumlin fields, dispersal trains, and ice streams in Arctic Canada. Canadian Geographer, **32:** 86-90.

Dyke, A. S., Morris, T. F., and Green, D. E. C. 1991. Postglacial tectonic and sea level history of the central Canadian Arctic. Geological Survey of Canada, Bulletin 397.

Dyke, A.S. *et al.* in press. Quaternary geology of Prince of Wales Island, Arctic Canada. Geological Survey of Canada, Memoir 433.

Edlund, S. A., and Alt, B. T. 1989. Regional congruence of vegetation and summer climate patterns in the Queen Elizabeth Islands, NWT. Arctic, **42:** 3-23.

Edlund, S. A., Alt, B. T., and Young, K. L. 1989. Interaction of climate, vegetation, and soil hydrology at Hot Weather Creek, Fosheim Peninsula, Ellesmere Island, NWT. Geological Survey of Canada, Paper **89-1D:** 125-133.

Edlund, S. A., and Woo, M. 1992. Eolian deposition on western Fosheim Peninsula, Ellesmere Island, Northwest Territories during the winter of 1990-91. Geological Survey of Canada, Paper **92-1B:** 91-96.

England, J. 1974. Advance of the Greenland Ice Sheet on to northeastern Ellesmere Island. Nature, **252:** 373-375.

England, J. 1976*a*. Late Quaternary glaciation of the eastern Queen Elizabeth Islands, NWT, Canada: alternative models. Quaternary Research, **6:** 185-202.

England, J. 1976*b*. Postglacial isobases and uplift curves from the Canadian and Greenland High Arctic. Arctic and Alpine Research, **8:** 61-78.

England, J. 1978. The glacial geology of northeastern Ellesmere Island, NWT, Canada. Canadian Journal of Earth Sciences, **15:** 603-617.

England, J. 1983. Isostatic adjustments in a full glacial sea. Canadian Journal of Earth Sciences, **20:** 895-917.

England, J. 1985. The late Quaternary history of Hall Land, northwest Greenland. Canadian Journal of Earth Sciences, **22:** 1394-1408.

England, J. 1986. A paleoglaciation level for north central Ellesmere Island, NWT, Canada. Arctic and Alpine Research, **18:** 217-222.

England, J. 1987. Glaciation and the evolution of the Canadian high arctic landscape. Geology, **15:** 419-424.

England, J. 1990. The late Quaternary history of Greely Fiord and its tributaries, west-central Ellesmere Island. Canadian Journal of Earth Sciences, **27:** 255-270.

England, J. 1992. Postglacial emergence in the Canadian High Arctic: integrating glacioisostasy, eustasy, and late deglaciation. Canadian Journal of Earth Sciences, **29**.

England, J., and Bednarski, J. 1986. Postglacial isobases from northern Ellesmere Island and Greenland: new data. Géographie Physique et Quaternaire, **XL:** 299-305.

England, J., and Bradley, R. S. 1978. Past glacial activity in the Canadian High Arctic. Science, **200:** 265-270.

England, J., Bradley, R. S., and Miller, G. H. 1978. Former ice shelves in the Canadian High Arctic. Journal of Glaciology, **20:** 393-404.

England, J., Bradley, R. S., and Stuckenrath, R. 1981. Multiple glaciations and marine transgressions, western Kennedy Channel, NWT, Canada. Boreas, **10:** 71-89.

England, J., Sharp, M., Lemmen, D. S., and Bednarski, J. 1991. On the extent and thickness of the Innuitian Ice Sheet: a postglacial-adjustment approach: Discussion. Canadian Journal of Earth Sciences, **28**: 1689-1695.

Evans, D. J. A. 1988. Glacial geomorphology and Late Quaternary history of Phillips Inlet and the Wootton Peninsula, Northwest Ellesmere Island, Canada. Unpublished PhD thesis, University of Alberta, Edmonton, Alberta.

Evans, D. J. A. 1989. An early Holocene narwhal tusk from the Canadian High Arctic. Boreas, **18**: 43-50.

Evans, D. J. A. 1990. The last glaciation and relative sea level history of NW Ellesmere Island, Canadian High Arctic. Journal of Quaternary Science, **5**: 67-82.

Ewing, M. and Donn, W.L. 1956. A theory of ice ages. Science, **123**: 1061-1066.

Fahey, B. D. 1985. Salt weathering as a mechanism of rock breakup in cold climates: an experimental approach. Zeitschrifte fur Geomorphologie, **29**: 99-111.

Fairbanks, R. G. 1989. A 17000 year glacio-eustatic sea level record: influence of glacial melting rates on the Younger Dryas event and deep ocean circulation. Nature, **342**: 637-642.

Firestone, J., Waddington, E., and Cunningham 1990. The potential for basal melting under Summit, Greenland. Journal of Glaciology, **36**: 163-168.

Fisher, D. A., and Koerner, R. M. 1981. Some aspects of climate change in the High Arctic during the Holocene as deduced fron ice cores. In: *Quaternary Paleoclimate*, Mahaney, W. C., ed., Geobooks; Norwich, 249-271.

Fisher, D. A., Koerner, R. M., Paterson, W. S. B., Dansgaard, W., Grudenstrup, N., and Reeh, N. 1983. Effects of wind scouring on climatic records from ice-core oxygen-isotope profiles. Nature, **301**: 205-209.

Flint, R. F. 1971. Glacial and Quaternary geology. John Wiley and Sons; New York, 872 p.

Fortier, Y. O., Blackadar, R. G., Glenister, B. F., Greiner, H. R., McLaren, D. J., McMillan, N. J., Norris, A. W., Roots, E. F., Souther, J. G., Thorsteinsson, R., and Tozer, E. T. 1963. Geology of north-central part of the Arctic Archipelago, Northwest Territories (Operation Franklin). Geological Survey of Canada, Memoir **320**: 617 p.

Fortier, Y. O., and Morley, L. W. 1956. Geological unity of the Arctic Islands. Transactions of the Royal Society of Canada, **L**: 3-12.

Francis, J. E. 1988. A 50 million-year-old fossil forest from Strathcona Fiord, Ellesmere Island, Arctic Canada: Evidence for a warm polar climate. Arctic, **41**: 314-318.

Francis, J. E. 1991. The dynamics of polar fossil forests: Tertiary fossil forests of Axel Heiberg Island, Canadain Arctic Archipelago. Geological Survey of Canada, Bulletin **403**: 29-38.

French, H. M. 1976. The Periglacial Environment. Longmans; London, 309p.

French, H. M., and Harry, D. G. 1990. Observations on buried glacier ice and massive segragated ice, western Arctic coast, Canada. Permafrost and Periglacial Precesses, **1:** 31-43.

Funder, S., Abrahamsen, N., Bennike, O., and Feyling-Hanssen, R. W. 1985. Forested Arctic: Evidence from North Greenland. Geology, **13:** 542-546.

Funder, S., and Houmark-Nielsen, M. 1990. Local events and regional glaciation. In: *Late Quaternary stratigraphy and glaciology in the Thule area, northwest Greenland*, Funder, S., ed., Meddelelser om Gronland, 22; 40-43.

Funder, S., Petersen, K. S., and Simonarson, L. A. 1987. The Early Pleistocene Arctic Ocean: view from the beach. Program with Abstracts, International Union for Quaternary Research, International Congress 12, p. 170.

Fyles, J. G. 1962. Surfical geology, Axel Heiberg-Ellesmere Island, 1961. In: *Field Work, 1961; Geological Survey of Canada, Information Circular no. 5*, Jenness, S. E., ed., 4-6.

Fyles, J. G. 1989. High terrace sediments, probably of Neogene age, west-central Ellesmere Island, NWT. Geological Survey of Canada, Paper **89-1D:** 101-104.

Fyles, J. G. 1990. Beaufort Formation (late Tertiary) as seen from Prince Patrick Island, Arctic Canada. Arctic, **43:** 393-403.

Fyles, J. G., Marincovich, L., Matthews, J. V., Jr., and Barendregt, R. 1991. Unique mollusc find in the Beaufort Formation (Pliocene) on Meighen Island, Arctic Canada. Geological Survey of Canada, Paper **91-1D:** 105-112.

Gilbert, M. W., and Clark, D. L. 1983. Central Arctic Ocean paleooceanographic interpretations based on Late Cenozoic calcareous dinoflagellates. Marine Micropaleontology, **7:** 385-401.

Gray, J. M. 1983. The measurement of shoreline altitudes in areas affected by glacio-isostasy, with particular reference to Scotland. In: *Shorelines and Isostasy*, Smith, D. E.; Dawson, A. G., eds., Insistute of British Geographers, Special Publication No. 16; London, 97-128.

Hart, J. K., and Boulton, G. S. 1991. The interrelation of glaciotectonic and glaciodepositional processes within the glacial environment. Quaternary Science Reviews, **10:** 335-350.

Hattersley-Smith, G. 1969. Glacial features of Tanquary Fiord and adjoining areas of northern Ellesmere Island, N.W.T. Journal of Glaciology, **8:** 23-50.

Hattersley-Smith, G. 1972. Climate change and related problems in northern Ellesmere Island, N.W.T. Canada. In: *Climate changes in Arctic areas during the last ten thousand years*, Yasari, Y, ed, University of Oulu, Oulu, Finland, 137-147.

Hebbelm, D., and Wefer, G. 1991. Effects of ice coverage and ice rafted material on sedimentation in the Fram Strait. Nature, **350:** 409-411.

Hein, F. J., and Mudie, P. J. 1991. Glacial-marine sedimentation, Canadian Polar Margin, north of Axel Heiberg Island. Géographie Physique et Quaternaire, **45:** 213-227.

Herman, Y. 1989. The Arctic Seas: climatology, oceanography, geology and biology. Van Nostrand Reinhold Co.; New York.

Herman, Y., Osmond, J. K., and Sonayajuly, B. L. K. 1989. Late Neogene arctic paleoceanography: Micropaleontology, stable isotopes, and chronology. In: *The arctic seas: Climatology, oceanography, geology and biology*, Herman, Y., ed., Van Nostrand Reinhold Company; New York, 581-656.

Herman, Y., and Hopkins, D. M. 1980. Arctic Ocean Climate in Late Cenozoic Time. Science, **209:** 557-562.

Hills, L. V. 1975. Late Tertiary floras of arctic Canada: An interpretation. In: *Circumpoloar Conference on Northern Ecology*, National Research Council of Canada, 165-171.

Hills, L. V., and Matthews Jr., J. V. 1974. A preliminary list of fossil plants from the Beaufort Formation, Meighen Island, District of Franklin. Geological Survey of Canada, Paper **74-1B:** 224-226.

Hills, L. V., and Ogilvie, R. T. 1970. *Picea banksii n.* sp., Beaufort Formation (Tertiary), Banks Island, Arctic Canada. Canadian Journal of Botany, **48:** 457-464.

Hodgson, D. A. 1974. Surficial geology, geomorphology and terrain disturbance, central Ellesmere Island, District of Franklin (Parts of 49D,E,G,H 340B). Geological Survey of Canada, Paper **74-1A:** 87-88.

Hodgson, D. A. 1985. The last glaciation of west-central Ellesmere Island, Arctic Archipelago, Canada. Canadian Journal of Earth Sciences, **22:** 347-368.

Hodgson, D. A. 1989. Quaternary geology of the Queen Elizabeth Islands. In: *Quaternary geology of Canada, Greenland*, Fulton, R. J., ed., Geological Survey of Canada, Geology of Canada, no. 1, 441-459.

Hodgson, D. A. 1990. Were erratics moved by glaciers or icebergs to Prince Patrick Island, western Arctic Archipelago, NWT. Geological Survey of Canada, Paper **90-1D:** 67-70.

Hodgson, D. A., and Edlund, S. A. 1977. Biophysical Regions, western Fosheim Peninsula and eastern Axel Heiberg Island. Geological Survey of Canada, Open File 501.

Hodgson, D. A., St-Onge, D. A., and Edlund, S. A. 1991. Surficial materials of Hot Weather Creek basin, Ellesmere Island, Northwest Territories. Geological Survey of Canada, Paper **91-1E:** 157-163.

Horn, D. R. 1963. Marine geology, Peary Channel, District of Franklin. Geological Survey of Canada, Paper **63-11:** 33p.

Hudec, P. P. 1973. Weathering of rocks in arctic and subarctic environments. Proceedings of the Symposium on the Geology of the Canadian Arctic, Saskatoon, 1973. J.D. Aitken and D.J. Glass (eds), 313-335.

Hughes, T. J., Denton, G. H., and Fastook, J. L. 1985. The Antarctic ice sheet: an analog for northern hemisphere paleo-ice sheets? In: *Models in Geomorphology*, Woldenberg, M. J., ed., Allen and Unwin; London, 25-72.

Hughes, T. J., Denton, G. H., and Grosswald, M. G. 1977. Was there a late Würm Arctic Ice Sheet? Nature **266:** 595-602.

Imbrie, J., and Imbrie, J. Z. 1980. Modelling the climatic response to orbital variations. Science, **202:** 943-953.

Irving, E., and Wynne, P. J. 1991. The paleolatitude of the Eocene fossils forests of Arctic Canada. Geological Survey of Canada, Bulletin **403:** 209-211.

Jansen, E., Bleil, U., Henrich, R., and Kringstad, L. 1988. Paleoenvironmental changes in the Norwegian Sea and the Northeast Atlantic during the last 2.8 Ma: Deep Sea Drilling Project/Ocean Drilling Program Sites 610, 642, 643, and 644. Paleooceanography, **3:** 563-581.

Jansen, E., and Sjøholm, J. 1991. Reconstruction of glaciation over the past 6 Ma from ice-borne deposits in the Norwegian Sea. Nature, **349:** 600-603.

Johnson, J. P., Jr. 1985. Geomorphic evidence and the upper marine limit in northern Labrador. In: *Our Geographic Mosaic: Research essays in honour of G.C. Mennill*, Knight, D. B., ed., Carlton University Press; Ottawa, 70-91.

Kaufman, D. S. 1992. Aminostratigraphy of Pliocene-Pleistocene high-sea-level deposits, Nome coastal plain and adjacent nearshore area Alaska. Geological Society of America Bulletin, **104:** 40-52.

Kaufman, D. S., Farmer, G. L., Miller, G. H., Carter, L. D., and Brigham-Grette, J. 1990. Strontium isotopic dating of upper Cenozoic marine deposits, northwestern Alaska. In: *Pliocene climates: Scenario for global warming. Abstracts from USGS Workshop, Denver, 23-25 October 1989*, Gosnell, L. B.; Poore, R. Z., eds., U.S. Geological Survey Open-File Report 90-64, 17-21.

Kaufman, D. S., Hopkins, D. M., Brigham-Grette, J., and Miller, G. H. 1989. Amino acid geochronology of marine and glacial events at Nome, Alaska. Geological Society of America Abstracts with Programs, **21:** A210.

Kerr, J. W. 1980. Structural framework of Lancaster Aulucogen, Arctic Canada. Geological Survey of Canada, Bulletin 319.

Kerr, J. W. 1981. Evolution of the Canadian Arctic Islands: a transition between the Atlantic and Arctic Oceans. In: *The Ocean Basin Margin, Volume 5, The Arctic Ocean*, Nairn, A. E. M.; Churkin, M., Jr.; Stehli, F. G., eds., Plenum Press; New York, 115-199.

Kerr, J. W., and De Vries, D. D. S. 1977. Structural geology of Somerset Island and Boothia Peninsula. Geological Survey of Canada, Paper **77-1A:** 107-111.

King, L. 1983. Contribution to the glacial history of the Borup Fiord area, northern Ellesmere Island, N.W.T., Canada. In: *Late and Postglacial Oscillations of Glaciers*, Schroeder-Lanz, H., ed., A.A. Balkema; Rotterdam, 305-323.

Koerner, R. M. 1968. Fabric analysis of a core from the Meighen Ice Cap, Northwest Territories, Canada. Journal of Glaciology, **10:** 421-430.

Koerner, R. M. 1977. Ice thickness measurements and their implications with respect to past and present ice volumes in the Canadian High Arctic ice caps. Canadian Journal of Earth Sciences, **14:** 2697-2705.

Koerner, R. M., and Fisher, D. A. 1990. A record of Holocene summer climate from a Canadian high arctic ice core. Nature, **343:** 630-631.

Koerner, R. M., Fisher, D. A., and Paterson, W. S. B. 1987. Wisconsinan and pre Wisconsinan ice thicknesses on Ellesmere Island, Canada: inferences from ice cores. Canadian Journal of Earth Sciences, **24:** 296-301.

Lemmen, D. S. 1988. The glacial history of Marvin Peninsula, northern Ellesmere Island, and Ward Hunt Island, high arctic Canada. Ph.D. Thesis. University of Alberta, Edmonton, Alberta.

Lemmen, D. S. 1989. The last glaciation of Marvin Peninsula, northern Ellesmere Island, Canada. Canadian Journal of Earth Sciences, **26:** 2578-2590.

Lemmen, D. S. 1990. Glaciomarine sedimentation in Disraeli Fiord, High Arctic Canada. Marine Geology, **94:** 9-22.

Lemmen, D. S., and England, J. 1992. Multiple glaciations and sea level changes, northern Ellesmere Island, high arctic Canada. Boreas, **21:** 1-16.

Lemmen, D. S., Evans, D. J. A., and England, J. 1988. Ice shelves of northern Ellesmere Island, NWT. Canadian Geographer, **32:** 363-367.

Lemmen, D. S., Gilbert, R., and Aitken, A. E. 1991. Quaternary investigations in the Expedition Fiord area, west-central Axel Heiberg Island, NWT. Geological Survey of Canada, Paper **91-1B:** 1-7.

Linick, T. W., Damon, P. E., Donahue, D. J., and Jull, A. J. T. 1989. Atomic Mass Spectrometry: the new revolution in radiocarbon dating. Quaternary International, **1:** 1-6.

Long, A. 1967. Age of trapped seawater at the bottom of Lake Tubourg, Ellesmere Island, N.W.T. Transactions of the American Geophysical Union, **48:** 136.

Lowdon, J. A., and Blake, W., Jr. 1979. Geological Survey of Canada Radiocarbon Dates XIX. Geological Survey of Canada, Paper **79-7:** 58p.

Lyons, J. B., and Mielke, J. E. 1973. Holocene history of a portion of Northernmost Ellesmere Island. Arctic, **26:** 314-323.

Mackiewicz, N. E., Powell, R. D., Carlson, P. R., and Molnia, B. F. 1984. Interlaminated ice-proximal glacimarine sediments in Muir Inlet, Alaska. Marine Geology, **57:** 113-147.

MacLean, B., Sonnichsen, G., Vilks, G., Powell, C., Moran, K., Jennings, A., Hodgson, D., and Deonarine, B. 1989. Marine geological and geotechnical investigations in Wellington and adjacent channels, Canadian Arctic Archipelago. Geological Survey of Canada, Paper 89-11.

Marlowe, J. I. 1968. Sedimentology of the Prince Gustaf Sea area, District of Franklin. Geological Survey of Canada, Paper **66-29**· 314-323.

Matthews, J. V., Jr. 1976. Insect fossils from the Beaufort Formation: geological, biological significance. Geological Survey of Canada, Paper **76-1B:** 217-227.

Matthews, J. V., Jr., and Ovenden, L. E. 1990. Late Tertiary plant macrofossils from localities in northern North America (Alaska, Yukon, and Northwest Territories). Arctic, **43:** 364-392.

Maxwell, J. B. 1981. Climate regions of the Canadian Arctic Islands. Arctic, **34:** 225-240.

Mayewski, P. A., Denton, G. H., and Hughes, T. J. 1981. Late Wisconsinian ice sheets of North America. In: *The Last Great Ice Sheets*, Denton, G. H.; Hughes, T. J., eds., John Wiley; New York, 67-178.

McKenna, M. 1980. Eocene paleolatitude, climate, and mammals of Ellesmere Island. Palaeogeography, Palaeoclimatology, Palaeoecology, **30:** 349-362.

McMillan, N. J. 1963. Slidre River. In: *Geology of the North-central part of the Arctic Archipelago, Northwest Territories (Operation Franklin)*, Fortier, Y. O., ed., Geological Survey of Canada, Memoir 320, 400-407.

McMillan, N. J. 1986. Tertiary fossil forests in the Arctic (News Report). Episodes, **9:** 169-170.

McNeil, D. H. 1990. Tertiary marine events of the Beaufort-Mackenzie Basin and correlation of Oligocene to Pliocene marine outcrops in Arctic North America. Arctic, **43:** 301-313.

Miall, A. D. 1986. The Eureka Sound Group (Upper Cretaceous-Oligocene), Canadian Arctic Islands. Bulletin of Canadian Petroleum Geology, **34:** 240-270.

Miall, A. D. 1988. The Eurka Sound Group: alternative interpretations of the stratigraphy and paleogeographic evolution - discussion. Geological Survey of Canada, Paper **88-1D:** 143-147.

Miller, G. H., Bradley, R. S., and Andrews, J. T. 1975. The glaciation level and lowest equilibrium line altitude in the High Canadaian Arctic: maps, climate interpretation. Arctic and Alpine Research, **7:** 155-168.

Miller, G. H., and Brighen-Grette, J. 1989. Amino acid geochronology: resolution and precision in carbonate fossils. Quaternary International, **1:** 111-128.

Miller, G. H., and De Vernal, A. 1992. Will greenhouse warming lead to Northern Hemisphere ice-sheet growth? Nature, **355:** 244-246.

Mörner, N. 1987. Models of global sea level changes. In: *Sea-level Changes*, Tooley, M. J.; Shennan, I., eds., Basil Blackwell; London, 332-355.

Müller, F. 1964. Evidence of climate fluctuations on Axel Heiberg Island, Canadian Arctic Archipelago. In: *Proceedings of the Symposium on the Arctic Heat Budget and Atmosphere Circulation*, Fletcher, J. O., ed., Rand Corporation; Santa Monica, California, 136-156.

Newman, M. A. 1978. Narwhal. In: *Marine Mammals*, Haley, D., ed., Pacific Search Press, 139-144.

Paterson, W. S. B. 1977. Extent of the late Wisconsin glaciation in northwest Greenland and northern Ellesmere Island. Quaternary Research, **8:** 180-190.

Paterson, W. S. B., Koerner, R. M., Fisher, D. A., Johnsen, S. J., Clausen, H. R., Dansgaard, W., Bucher, P., and Oschger, H. 1977. An oxygen isotope climatic record from the Devon Island Ice Cap, Arctic Canada. Nature, **266:** 508-511.

Pelletier, B. R. 1966. Development of submarine physiography in the Canadian Arctic and its relation to crustal deformation. In: *Continental Drift*, Grailand, G. D., ed., Royal Society of Canada, Special Publication #9, 77-101.

Pollard, W. H. 1991. Observations on massive ground ice on Fosheim Peninsula, Ellesmere Island, Northwest Teritories. Geological Survey of Canada, Paper **91-1E:** 223-231.

Potts, A. S. 1970. Frost action in rocks: some experimental data. Transactions of the Institute of British Geographers, **49:** 109-124.

Prest, V. K. 1990. Laurentide ice-flow patterns: a historical review, and implications of the dispersal of Belcher Island erratics. Géographie Physique et Quaternaire, **44:** 113-136.

Raymo, M. E., Rind, D., and Ruddiman, W. F. 1990. Climatic effects of reduced arctic sea ice limits in the GISS II general circulation model. Paleoceanography, **5:** 367-382.

Reeh, N. 1984. Reconstruction of the glacial ice cover of Greenland and the Canadian Arctic Islands by three-dimensional perfectly plastic ice sheet modelling. Annals of Glaciology, **5:** 115-128.

Repenning, C. A. 1984. Quaternary rodent biochronology and its correlation with climatc and magnetic stratigraphies. In: *Correlation of Quaternary stratigraphies*, Mahaney, W. C., ed., Geobooks; Norwich, 105-118.

Repenning, C. A. 1990. Of mice and ice in the late Pliocene of North America. Arctic, **43:** 314-323.

Repenning, C. A., Brouwers, E. M., Carter, L. D., Marincovich, L., Jr., and Ager, T. A. 1987. The Beringian ancestry of *Phenaconys* and the beginning of the modern Arctic Ocean Borderland Biota. USGS Bulletin 1687.

Retelle, M. J. 1986. Glacial geology and Quaternary marine stratigraphy of the Robeson Channel area, northeastern Ellesmere Island, NWT. Canadian Journal of Earth Sciences, **23:** 1001-1012.

Retelle, M. J., Bradley, R. S., and Stuckenrath, R. 1989. Relative sea level chronology determined from raised marine sediments and coastal isolated basins, Northeast Ellesmere Island, Arctic Canada. Arctic and Alpine Research, **21:** 113-125.

Ricketts, B. D. 1987*a*. Preliminary structural cross-sections across Fosheim Peninsula and Axel Heiberg Island, Arctic Archipelago. Geological Survey of Canada, Paper **87-1A:** 369-374.

Ricketts, B. D. 1987*b*. Evolution of the eastern Arctic landscape and the Tertiary - Quaternary transition, Queen Elizabeth Islands. In: *Programme with Abstracts*, International Union for Quaternary Research, XII International Congress, Ottawa, Ontario, p. 252.

Ricketts, B. D. 1987*c*. Princess Margaret Arch: Re-evaluation of an element of the Eurekan Orogen, Axel Heiberg Island, Arctic Archipelago. Canadian Journal of Earth Sciences, **24:** 2499-2505.

Ricketts, B. D. 1988. The Eureka Sound Group: alternative interpretations of the stratigraphy and paleogeographic evolution - Reply. Geological Survey of Canada, Paper **88-1D:** 149-152.

Ricketts, B. D. 1991. Sedimentation, Eurekan tectonism and the fossil forest succession on eastern Axel Heiberg Island, Canadian Arctic Archipelago. Geological Survey of Canada, Bulletin **403:** 1-27.

Ricketts, B. D., and Embry, A. F. 1984. Summary of geology and resource potential of coal deposits in the Canadian Arctic Archipelago. Bulletin of Canadian Petroleum Geology, **32:** 359-371.

Ricketts, B. D., and McIntyre, D. J. 1986. The Eureka Sound Group, eastern Axel Heiberg Island: New data on the Eurekan Orogeny. Geological Survey of Canada, Paper **87-1B:** 405-410.

Rose, J. 1978. Glaciation and sea level change at Bugoyfjord, south Varangenfjord, North Norway. Norsk. Geografisk Tidsskrift, **32:** 121-135.

Rudberg, S. 1963. Geomorphological processes in a cold semi-arid region. In: *Preliminary Report 1961-1962, Axel Heiberg Research Reports*, McGill University; Montreal, 139-149.

Ruddiman, W. F., and Kutzbach, J. E. 1989. Forcing of the Cenozoic northern hemisphere climate by plateau uplift in southern Asia and the American West. Journal of Geophysical Research, **94:** 18409-18427.

Ruddiman, W. F., and Raymo, M. E. 1988. Northern Hemisphere climate regimes during the past 3 Ma: possible tectonic connections. Transactions of the Royal Society London, **B318:** 411-430.

Sagar, R. B. 1962. Meteorological and glaciological studies: ice rise station, Ward Hunt Island, May to September 1960. Arctic Institute of North America, Research Paper no. 24.

Sejrup, H. 1990. Amino acid geochronology. In: *Late Quaternary stratigraphy and glaciology in the Thule area, Northwest Greenland*, Funder, S., ed., Meddelelser om Gronland, 22, 36-39.

Shackleton, N. J., Backman, J., Zimmerman, H., Kent, D. V., Hall, M. A., Roberts, D. G., Schnitker, D., and Baldauf, J. 1984. Oxygen isotope calibration of the onset of ice-rafting and history of glaciation in the North Atlantic region. Nature, **307:** 620-623.

Shackleton, N. J., and Kennett, J. P. 1975. Paleotemerature history of the Cenozoic and the initiation of Antarctic glaciation: oxygen and carbon isotope analysis in DSDP sites 277, 279, 281. In: *Initial Reports of the Deep Sea Drilling Project, # 29*, United States Government Printing Office; Washington, D.C., 743-755.

Shackleton, N. J., and Opdyke, N. D. 1977. Oxygen isotope and paleomagnetic evidence for early Northern Hemisphere glaciation. Nature, **270**: 216-219.

Shaw, J. 1982. Melt-out till in the Edmonton area, Alberta, Canada. Canadian Journal of Earth Sciences, **19**: 1548-1569.

Sher, A. V., Kapina, T. N., Kouznetsov, Y. V., Virina, E. I., and Zazhigin 1979. Late Cenozoic of the Kolyma Lowland. In: *XIV Pacific Science Congress: Tour Guide XI*, Academy of Sciences of USSR, 155p.

Sim, V. W. 1961. A note on high-level marine shells on Fosheim Peninsula, N.W.T. Geographical Bulletin, **16**: 120-123.

Sloan, V. F. 1990. The glacial history of central Canon Fiord, west-central Ellesmere Island, Arctic Canada. M.Sc. thesis. University of Alberta, Edmonton, Alta.

Smith, N. D. 1985. Proglacial fluvial environment. In: *Glacial Sedimentary Environments, SEPM Short Course no. 16*, Ashley, G. M.; Shaw, J.; Smith, N. D., eds., Society of Pa;eontologists and Mineralogists; Tulsa, OK, 85-134.

Sobczak, L. W., Mayr, U., and Sweeney, J. F. 1986. Crustal section across the polar continent - ocean transition in Canada. Canadian Journal of Earth Sciences, **23**: 608-621.

Stewart, T. G. 1991. Glacial marine sediemntation from tidewater glaciers in the Canadian high arctic. In: *Glacial marine sedimentation: paleoclimatic significance*, Anderson, J. B.; Ashley, G. M., eds., GSA Special Paper 261, 95-106.

Stewart, T. G., and England, J. 1983. Holocene sea-ice variations and plaeoenvironmental change, northernmost Ellesmere Island, N.W.T., Canada. Arctic and Alpine Research, **15**: 1-17.

Sugden, D. E. 1977. Reconstruction of the morphology, dynamics, and thermal characteristics of the Laurentide Ice Sheet at its maximum. Arctic and Alpine Research, **9**: 21-47.

Sugden, D. E., and Watts, S. H. 1977. Tors, felsenmeer and glaciation in northern Cumberland Peninsula, Baffin Island. Canadian Journal of Earth Sciences, **14**: 2817-2823.

Sutherland, D. G. 1981. The raised shorelines and deglaciation of Loch Long/Loch Fyne area, western Scotland. Unpublished PhD thesis, University of Edinburgh, Edinburgh.

Sverdrup, O. 1904. New Lands, Four Years in the Arctic Regions Vol. 2. Longmans; London.

Tarnocai, C., and Smith, C. A. S. 1991. Paeosols of the fossil forest area, Axel Heiberg Island. Geological Survey of Canada, Bulletin **403**: 171-187.

Thorsteinsson, R., and Tozer, E. T. 1970. Geology of the Arctic Archipelago. In: *Geology and Economic Minerals of Canada*, Douglas, R. J. W., ed., Geology Survey of Canada, Economic Geology Report No. 1, 547-590.

Tozer, E. T. 1956. Geological reconnaissance, Price Patrick, Elington, and western Melville islands, Arctic Archipelago, Northwest Teritories. Geological Survey of Canada, Paper **55-5**: 32p.

Tozer, E. T., and Thorsteinsson, R. 1964. Western Queen Elizabeth Islands, Arctic Archipelago. Geological Survey of Canada, Memoir **332:** 242p.

Trettin, H. P. 1987. Pearya: a composite terrane with Caledonian affinities in northern Ellesmere Island. Canadian Journal of Earth Sciences, **24:** 224-245.

Trettin, H. P. 1989. The Arctic Islands. In: *The Geology of North America - An Overview*, Bally, A. W.; Palmer, A. R., eds., Geological Society of America, The Geology of North America, v. A.; Boulder, Colorado, 349-370.

Troelsen, J. C. 1952. Geological investigations in Ellesmere Island, 1952. Arctic, **5:** 199-210.

Tushingham, A. M. 1991. On the extent and thickness of the Innuitian Ice Sheet: a postglacial adjustment approach. Canadian Journal of Earth Sciences, **28:** 231-239.

Walcott, R. I. 1970. Isostatic response of loading of the crust in Canada. Canadian Journal of Earth Sciences, **7:** 716-726.

Watts, S. H. 1983. Weathering pit formation in bedrock near Cory glacier, southeastern Ellesmere Island, N.W.T. Geological Survey of Canada, Paper **83-1A:** 487-491.

Wehmiller, J. F. 1984. Relative and absolute dating of Quaternary molluscs with amino acid racemization: evaluation, application, and questions. In: *Quaternary Dating Methods*, Mahaney, W. C., ed., Elsevier; Atmersterdam, 171-193.

Wellman, H. W., and Wilson, A. T. 1965. Salt weathering: a neglected geological erosive agent in coastal and arid environments. Nature, **205:** 1097-1098.

West, R. M., Dawson, M. R., and Hutchison, J. H. 1977. Fossils from the Paleogene Eureka Sound Formation, NWT, Canada: occurrence, climatic and paleogeographic implications. In: *Paleontology and Plate Tectonics*, West, R. M., ed., Milwaukee Public Museum, Special Papers in Biology and Geology #2, 77-93.

White, W. A. 1972. Deep erosion by continental ice sheets. Geological Society of America Bulletin, **83:** 1037-1056.

White, W. A. 1988. More on deep glacial erosion by continental ice sheets and their tongues of distributary ice. Quaternary Research, **30:** 137-150.

Wolfe, J. A. 1978. A paleobotanocal interpretation of Tertiary climates in the northern Hemisphere. American Scientist, **66:** 694-703.

Wolfe, J. A. 1980. Tertiary climates and floristic relationships at high latitudes in the northern hemisphere. Palaeogeography, Palaeoclimatology, Palaeoecology, **30:** 313-323.

Woo, M., Edlund, S. A., and Young, K. L. 1991. Occurrence of early snow-free zones on Fosheim Peninsula, Ellesmere Island, Northwest Territories. Geological Survey of Canada, Paper **91-1B:** 9-14.

Woo, M., Young, K. L., and Edlund, S. A. 1990. 1989 observations of soil, vegetation and microclimate and effects on slope hydrology, Hot Weather Creek basin, Ellesmere Island, N.W.T. Geological Survey of Canada, Paper 90-1D: 85-93.

Young, R. R. 1991. Late Pleistocene geomorphology of the Hand Hills area, southern Alberta. Unpublished MSc thesis, University of Alberta, Edmonton, Alberta.

APPENDIX A

Air photographs

The following air photographs illustrate some of the landform and sediment associations *described in the text*. Only the major features are labelled. Where a landform appears more than once in a photograph, some examples are left unmarked to allow an unobstructed view. On the oblique photographs, scale is provided by distance to recognizable landmarks. A solid arrow (➚) is used as a pointer. Other symbols and abbreviations include:

⇀	Ice-marginal channels	***Tb***	till blanket
↗	proglacial channels	***Tv***	till veneer
◸	raised marine delta	***Mv***	marine veneer
⌣	Holocene marine limit	***Hts***	high terrace sediment
⁘	outwash	***Gt***	gravel terrace
░	glaciomarine sediment		

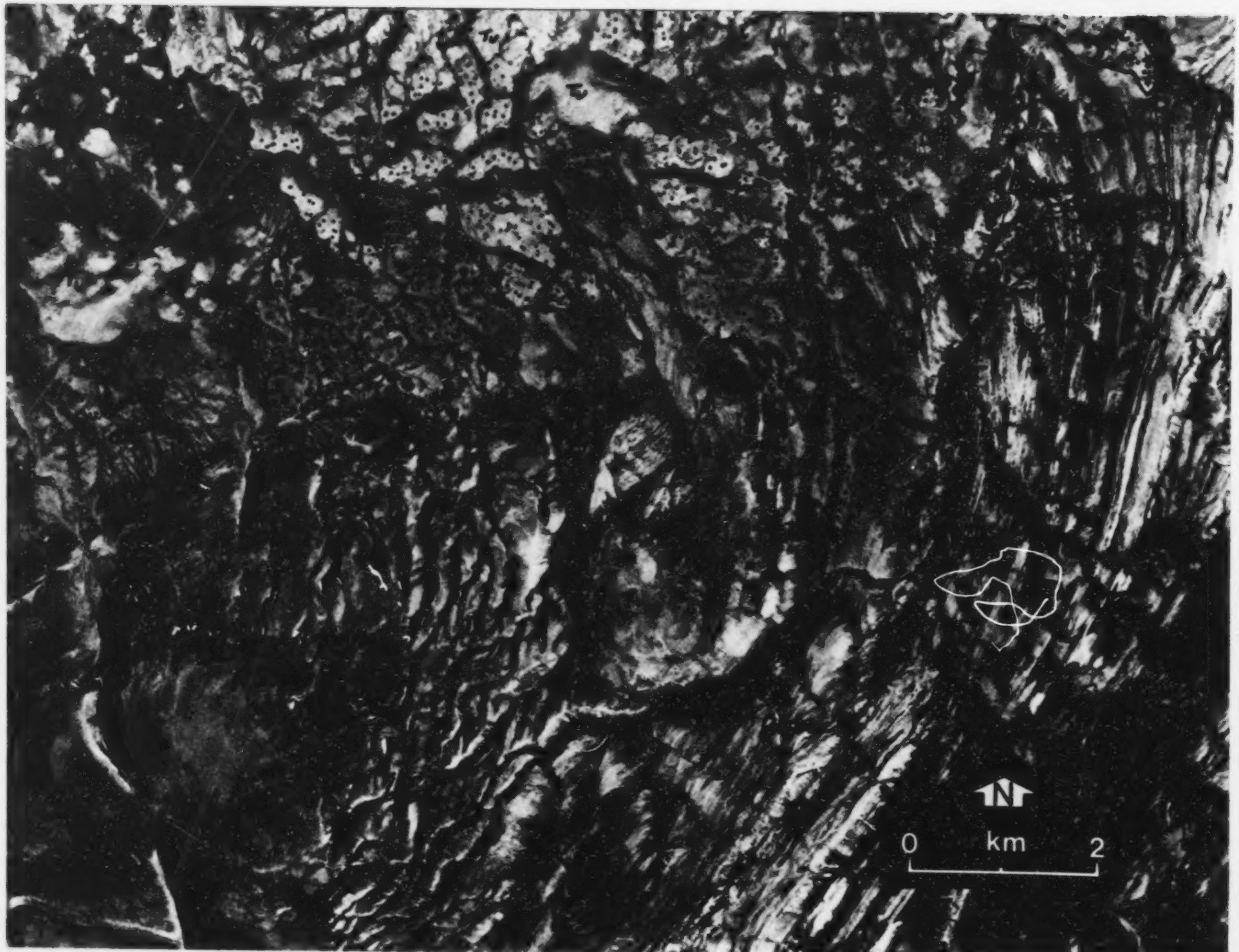


Figure A.1: Vertical air photograph of northern Black Top Ridge and adjacent lowland. National Air Photo Library (NAPL) #A16686-19.

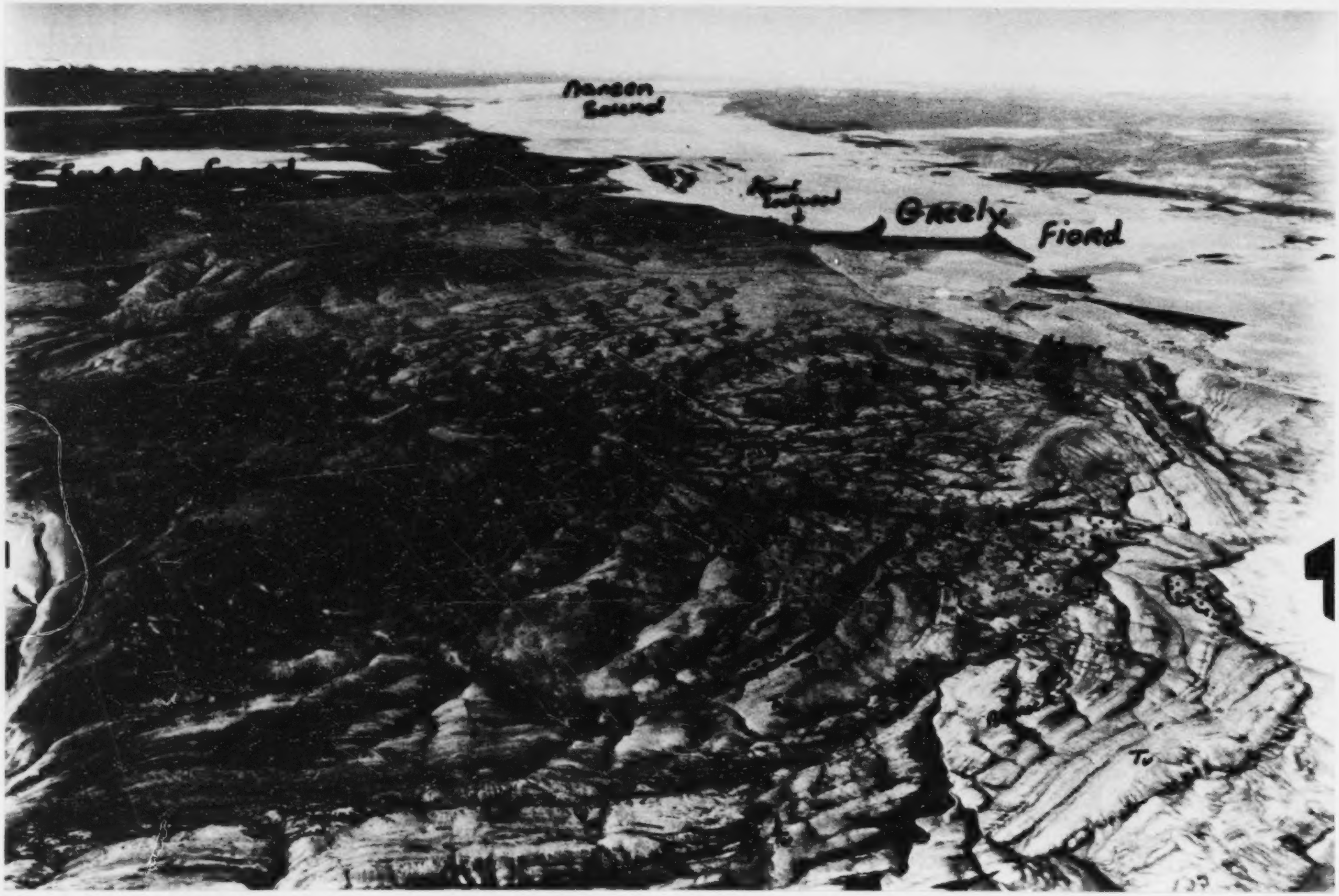

Nansen Sound
Eureka Sound
Mount Lockwood
Greely Fiord
Tv

Figure A.2: Oblique air photograph looking NW towards Nansen Sound. Black Top Ridge and Mount Lockwood are 5 and 25 km away, respectively. NAPL #T490L-9.

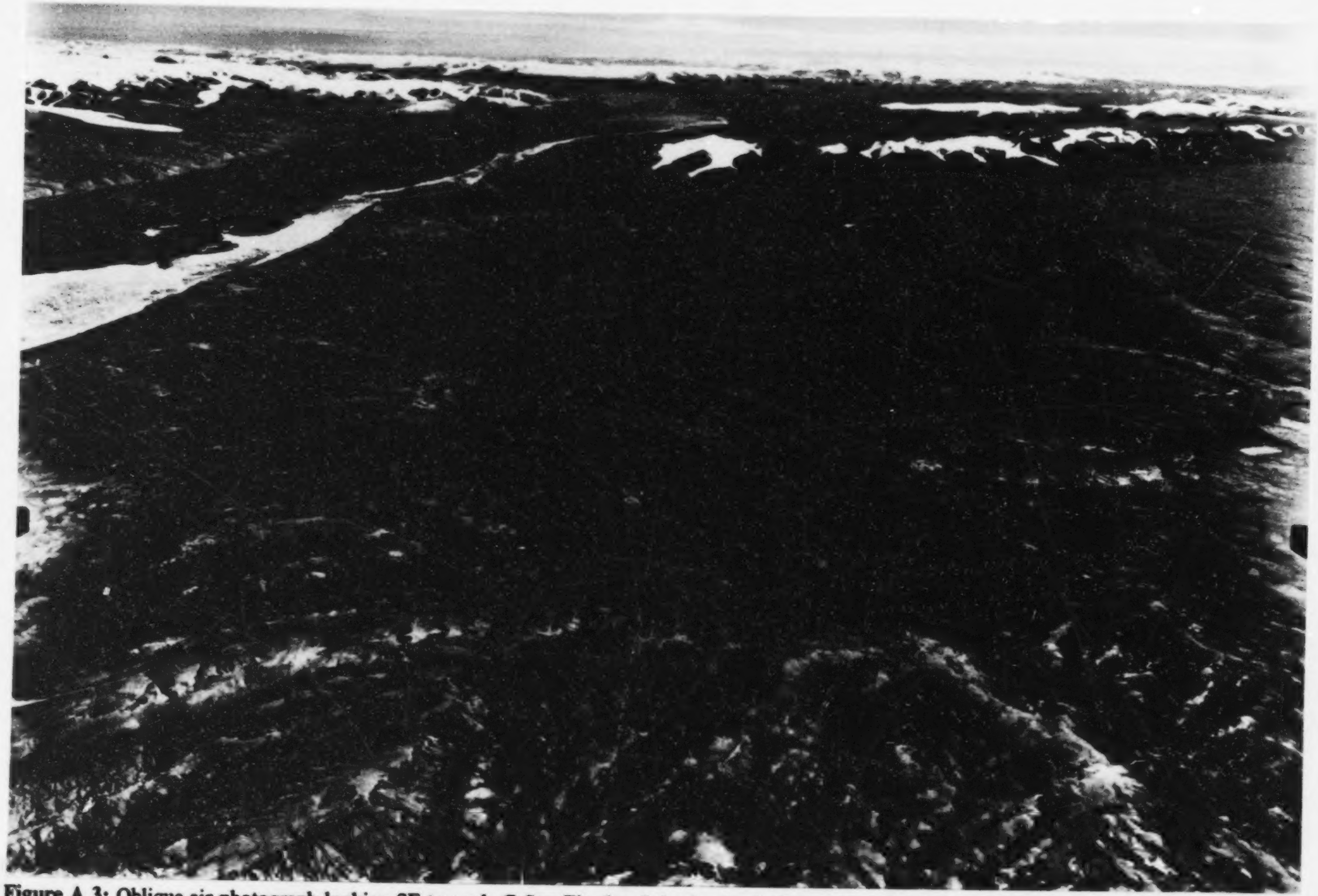


Figure A.3: Oblique air photograph looking SE towards Cañon Fiord and the Sawtooth Range. Mt. Bridgman is 28 km away. NAPL #T490R-11.

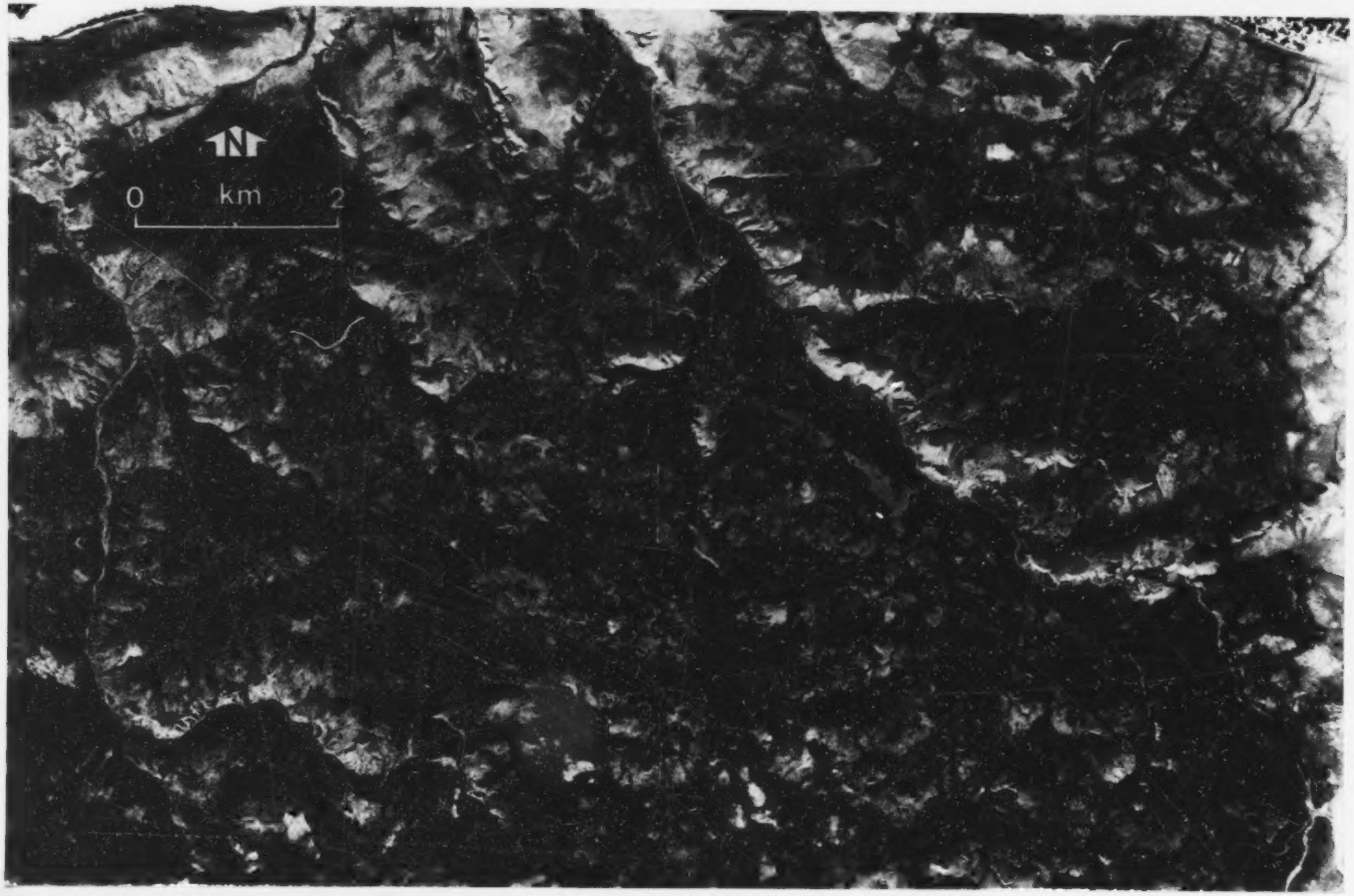


Figure A.4: Vertical air photograph of the coastal lowlands near Greely Fiord. NAPL #A16734-35.

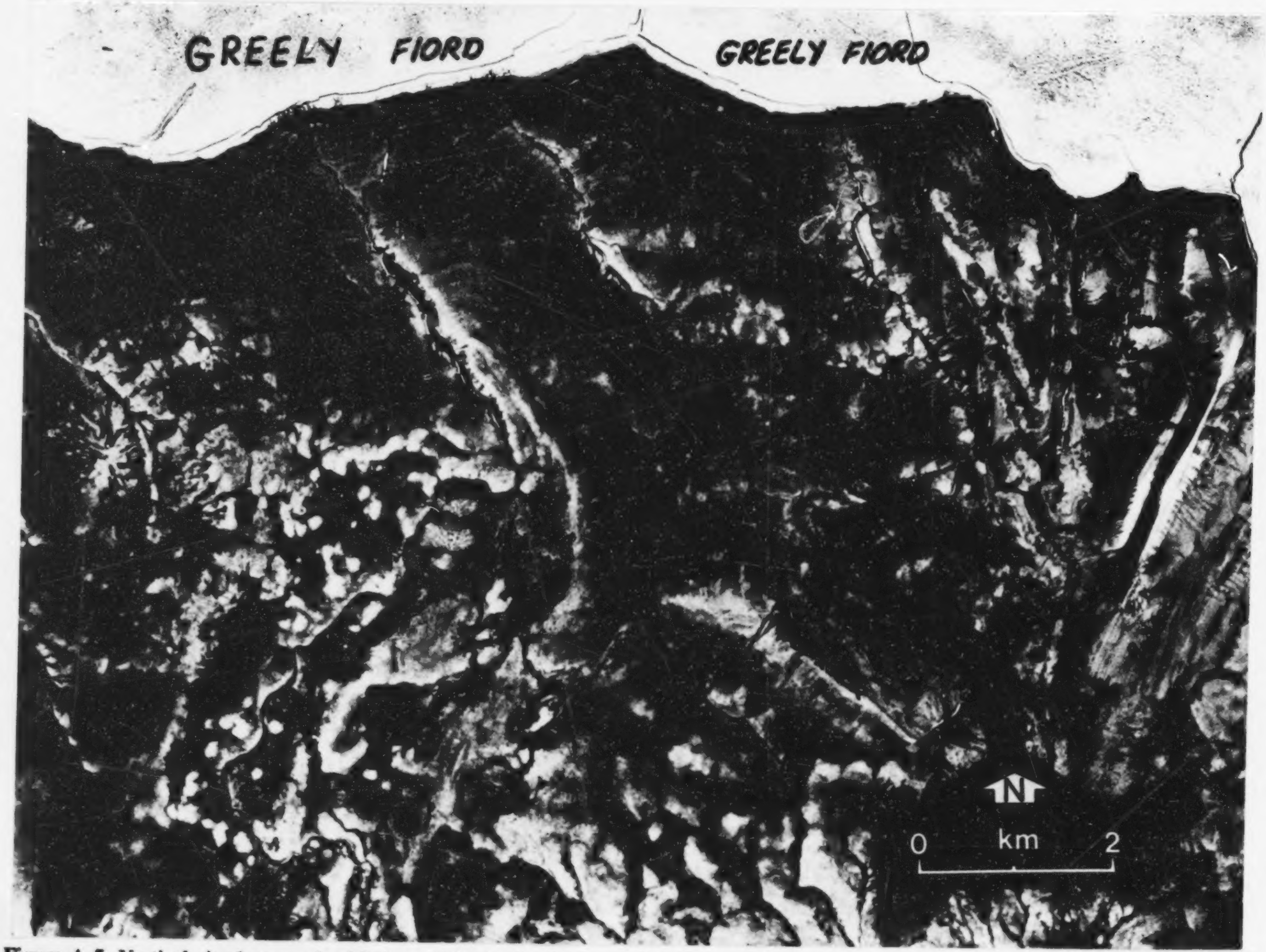

GREELY FIORD
GREELY FIORD
N
0 km 2

Figure A.5: Vertical air photograph of Iceberg Point lowland. Grounding lines and hummocky terrain on left. NAPL #A16605-50.

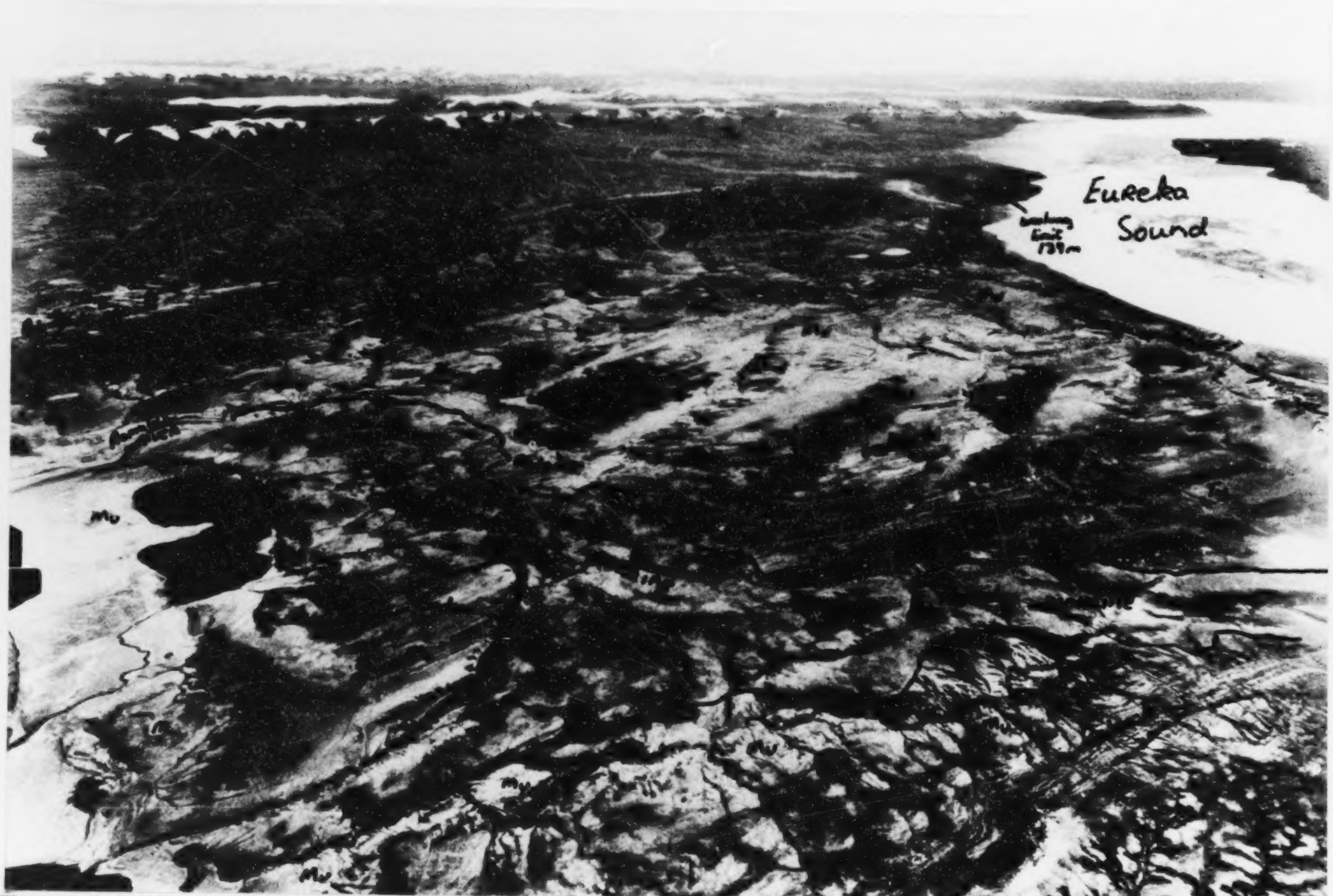

Eureka Sound
marine limit 139m

Figure A.6: Oblique air photograph looking SSE towards central Eureka Sound and the Sawtooth Range. Romulus Lake and Airfield ridge are 5 and 30 km away, respectively. NAPL #T492R-73.

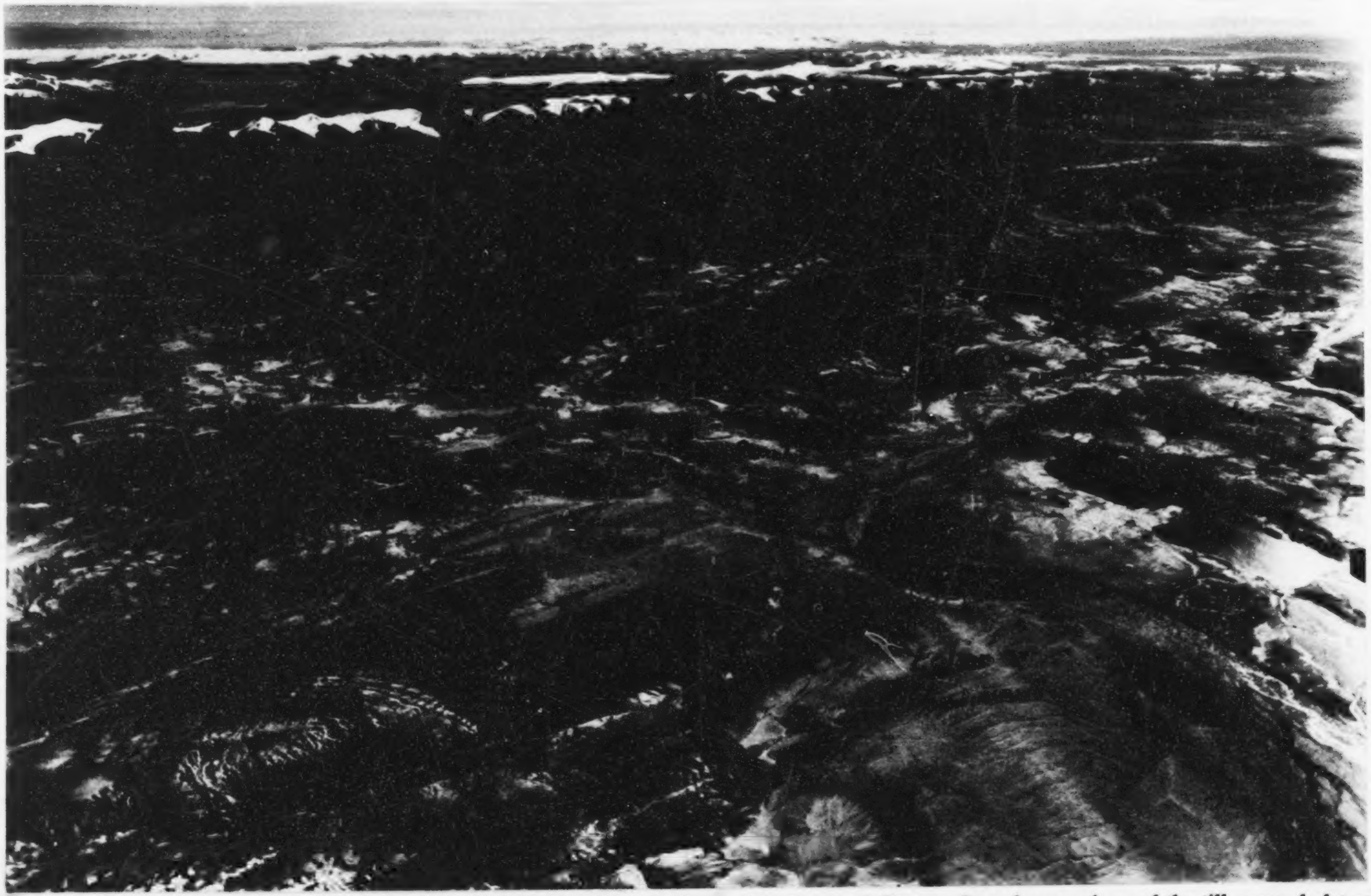


Figure A.7: Oblique air photograph looking SE towards Slidre River Valley and the Sawtooth Range. Romulus moraine and the till-covered plateau are 7 and 35 km away, respectively. NAPL #T490R-5.

APPENDIX B

Stratigraphic sections

The Quaternary stratigraphy of western Fosheim Peninsula is poorly exposed and sections are generally limited. The following is a general description of available sections and their provisional interpretation.

SECTION 1, CAPE LOCKWOOD

Near Cape Lockwood on the northeastern coast of Fosheim Peninsula, a deeply incised river valley has exposed 70 m of gently dipping beds of greyish sand at 130 m asl (top). Only the lower part of the section was accessible. Here, rhythmically laminated and thin planar-bedded, fine sand and sandy silt, with granule beds and coal fragments, were observed (Figs. B.1 and B.2). 1 to 2 m of gravelly diamicton unconformably overly the sand. Immediately beneath the diamicton, the sand exhibits pervasive deformation (folding and faulting) which decreases with depth. The till face was inaccessible; however, from the cliff top, I observed that boulders were striated and some retained a preferred long-axis orientation parallel to the valley, suggesting ice movement from the southeast. Pieces of unaltered wood lay at the base of the section.

Section 1 is tentatively interpreted to represent a sandar deposit, overlain by till. In many respects, the uniform texture and structure of the sand is atypical of glaciofluvial sediment which commonly displays a wide range of grain sizes and sedimentary structures (Smith 1985). However, the uniform texture of the sand may be a reflection of its source which is inferred to be the local Neogene terraces which contain thick sequences of sand and sandy silt (Fyles 1989; Hodgson *et al.* 1991). Unaltered wood in the section may have also originated from the Neogene terraces.

The gravelly diamicton capping the section is interpreted to be till because of its erratic composition (quartzite, sandstone, conglomerate), apparent fabric, striated clasts and lateral continuity. Deformation of the underlying sand is consistent with overriding by ice. A coarsening-upward sequence immediately below the till was not observed. The till cap precludes a Holocene age for the sandar and implies that the incision of Neogene terraces was under way prior to the last regional glaciation of the peninsula.

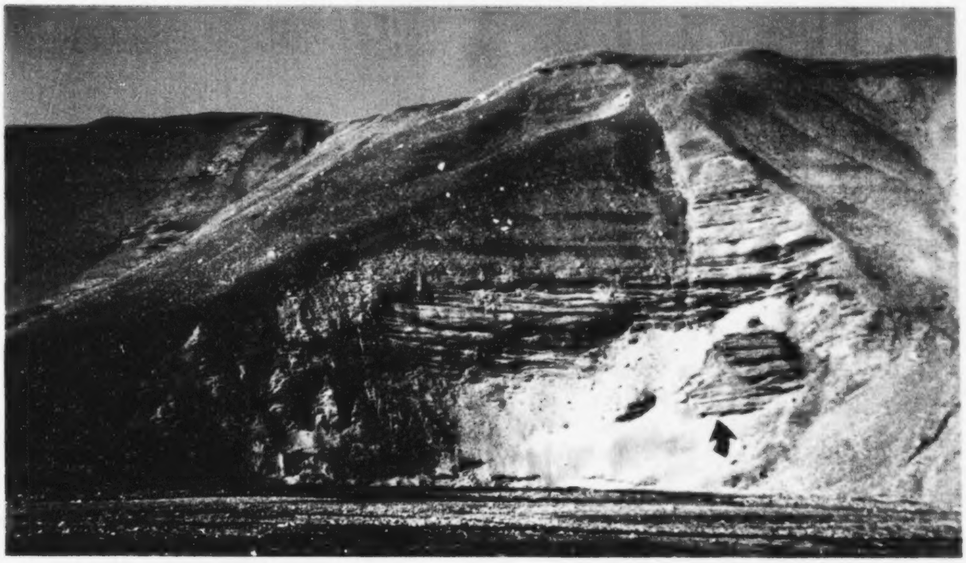


Figure B.1: Section 1 near Cape Lockwood exposes *ca.* 70 m of gently dipping sand overlain by a gravelly till cap (dashed line). Location of section in Figure B.2 is shown by arrow.

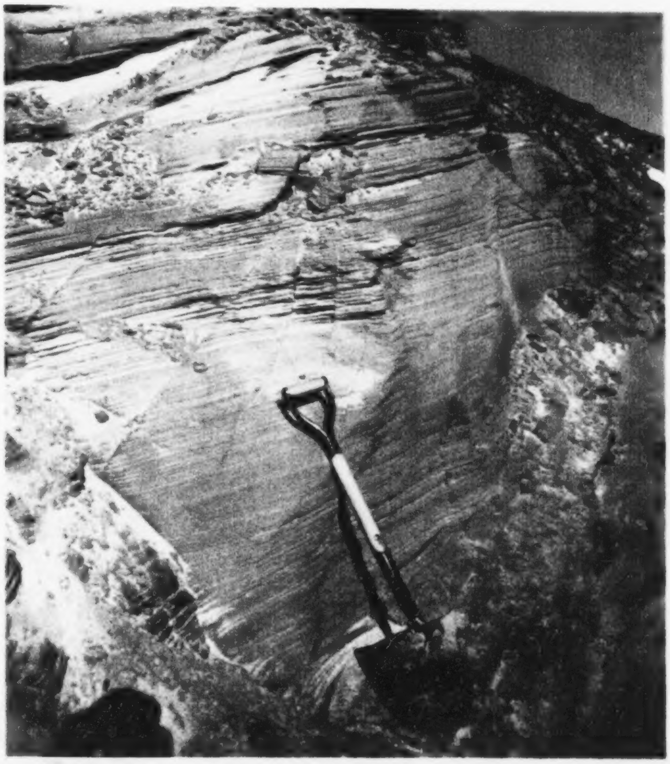


Figure B.2: Close-up of section 1 with rhythmically laminated fine sand and sandy silt with granule beds and coal fragments. Note normal faults (due to compaction?). Shovel is *ca.* 1 m long.

SECTION 2, SNOW GOOSE HILL

A limited section near Snow Goose hill exposed 1 m of fossiliferous diamicton which overlies, with angular unconformity, graded beds of sand and silt (Figs. B.3 and B.4). These beds exhibit pervasive shearing with silt injections along shear planes. The diamicton has a sharp, erosional contact and contains striated erratics and clasts of sub-diamicton sand.

The diamicton is interpreted as till deposited by meltout rather than lodgement because sand intraclasts are undeformed (*cf.* Shaw 1982). Postdepositional deformation is primarily due to extensional forces where the more cohesive sand behaved as a rigid body, failing along shear planes and the silt reacted plastically, having been injected along the shear planes. This extensional deformation is consistent with the high shear strain imposed by overriding ice (Hart and Boulton 1991).

SECTION 3, ICEBERG POINT

An excavated section revealed 50-75 cm of diamicton overlying interbedded fine to coarse sand with dispersed granules, small pebbles and shell fragments (sites 24 and 28; Figs. B.5 and B.6). The sand exhibits a decrease in deformation down from the contact with the overlying diamicton. The upper 10-15 cm of sand has been sheared forming boudins of coarser sand within finer, less competent sand. At its upper contact, the sand has been folded into the diamicton and, in places, attenuated into stringers or laminations. Below this zone, normal faulting represents weaker deformation.

The lower 10-15 cm of diamicton has a dark brown colour due to disseminated organics within a clayey matrix. The upper part has a harder, silty matrix with undeformed clasts of sub-diamicton sand. Granite, quartzite and carbonate erratics and shell fragments occur throughout the diamicton (site 28).

The origin of the interbedded fine to coarse sand is uncertain but it may represent a proximal, glaciomarine deposit with granule- and pebble-size dropstones. The shell fragments may also be glacial erratics. Deformation observed in the sand is closely associated with deposition of the overlying diamicton which is interpreted to be of glacial origin. The decrease in shear deformation down from the diamicton base is a characteristic feature of subglacial tectonism where strain rate decreases downward from the glacier sole (Banham 1977; Boulton 1979). Small-scale recumbent folding at the contact between the two units results in interfingering of diamicton and sand, indicating that both were subject to the same deformational process. The attenuation of sand laminations in a down-ice direction from these folds suggests shear deformation; however, deposition by mass flow processes cannot be ruled out.

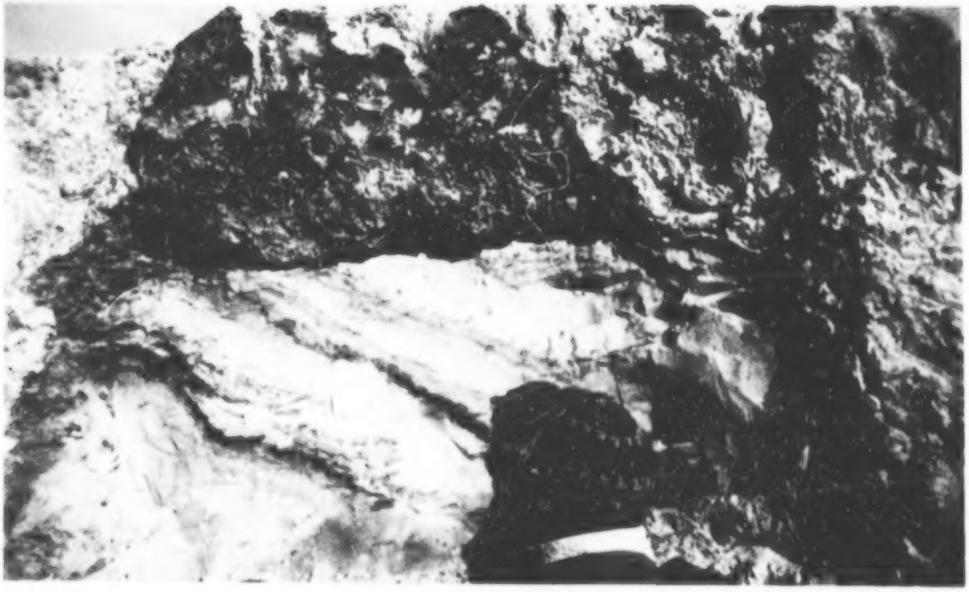


Figure B.3: Section 2 showing meltout till overlying glaciotectonized sand and silt. Undeformed sand clasts (arrows) occur within the till. Scale card is 9 cm wide.

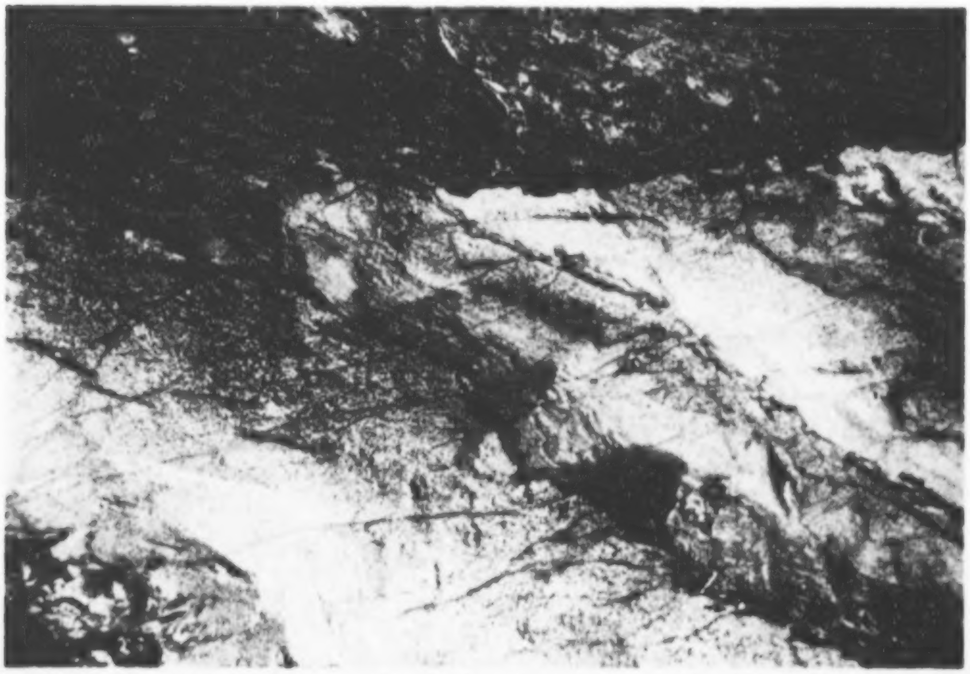


Figure B.4: View of the sharp, erosional basal till contact and the sheared and faulted sand in section 2. The shear planes are highlighted by silt injection. Displacement shown by arrows.

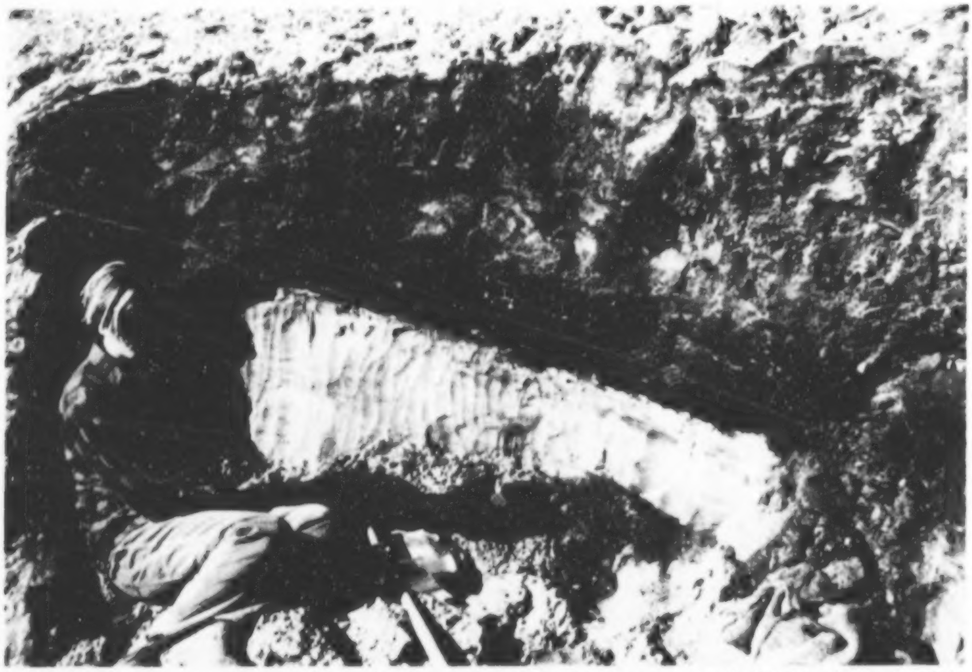


Figure B.5: Section 3 at 98 m asl near Iceberg Point. The section exposes glacial diamicton overlying subglacially deformed fine to coarse sand.

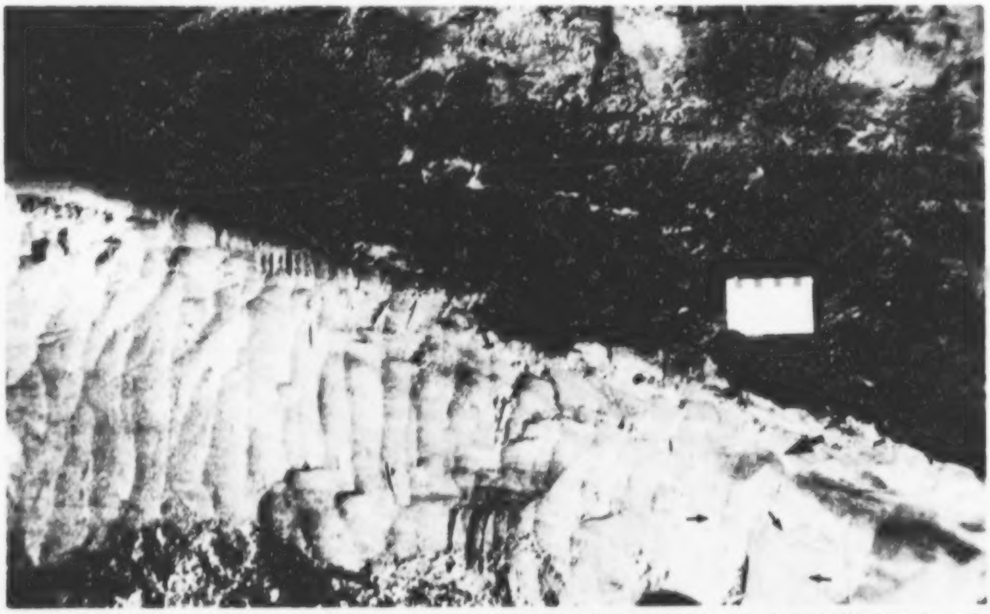


Figure B.6: View of till-sand contact in section 3. Note folding of sand along lower till contact and attenuation down-ice (to left). Sand boudins and normal faults are indicated on lower right.

SECTION 4, ROMULUS LOWLAND

A section along a tributary valley of Romulus river in the northern lowland revealed interbedded sand and diamicton (Figs. B.7 - B.12). The lowermost unit consists of laminated silty sand interbedded with climbing-ripple and cross-laminated sand. It is considered to represent periodic and rapid sedimentation. Paleocurrent indicators suggest flow from the southwest. A diamicton (*ca.* 2 m thick) rests on top of the sand with angular unconformity and contains striated clasts (no granites observed) which are generally oriented parallel to the inclined, basal contact. Deformation of the sub-diamicton units includes drag-folding and thrust faulting. These indicate emplacement of the diamicton from the southwest (Fig. B.8). An elongate raft of laminated sand occurs within the upper half of the diamicton and retains its original structure, although in some places it has been attenuated and deformed (Fig. B.8). This diamicton is interpreted as till deposited by ice advancing from Eureka Sound (southwest). It is not considered to be a debris flow because the underlying deformed sediments are displaced *upslope* along thrust planes. The relatively undeformed sand raft may have been incorporated into the ice by proglacial or subglacial thrusting and subsequently deposited by melt-out in the till.

The top of the till has a scour form which is infilled with several metres of cross laminated sand and horizontal, interbedded diamictons and finely laminated sand (Figs. B.9 and B.10). Paleocurrent indicators suggest flow from the east or northeast. The diamictons, ranging in thickness from 10 cm to 30 cm, contain pebble to cobble-size clasts and sand stringers. The lower contact of each diamicton is sharp but irregular with some folding of underlying sediment into the diamicton matrix. The finely laminated sand contains numerous reverse faults injected with silt (Fig. B.9).

This unit is interpreted as a sequence of debris flows interbedded with either channel or tunnel sand. The debris flows contain stringers of underlying sand and although their lower contacts are to some degree erosional, they also mimic the underlying sediment surface. (Note the drape of diamicton over the large coal fragment in Figure B.10). The sand may have been deposited in either a proglacial channel or a subglacial tunnel. A large boulder in the sand may have originated from a tunnel roof (*cf.* Shaw 1982). Till clasts were not observed in the sand as would be expected considering the channel/tunnel is eroded in till. This unit may represent either an ice-marginal, subaqueous environment or a subglacial environment.

Truncating these beds to the southwest (left, Fig. B.10) is an inclined diamicton which, during emplacement, caused drag-folding and thrust faulting in the underlying sediments. The inclined diamicton is at least 3-4 m thick and although some sand intrabeds appear to subdivide it into several beds, others pinch out to form sand lenses (Fig. B.10). This sand is similar to that found in the underlying unit and, apart from minor faults, is relatively undeformed. Clast orientation within the diamicton tends to mimic the slope of its basal contact. This diamicton is interpreted as till and not a debris flow because displacement of underlying sediments by thrusting and drag-folding is *upslope* towards the northeast.

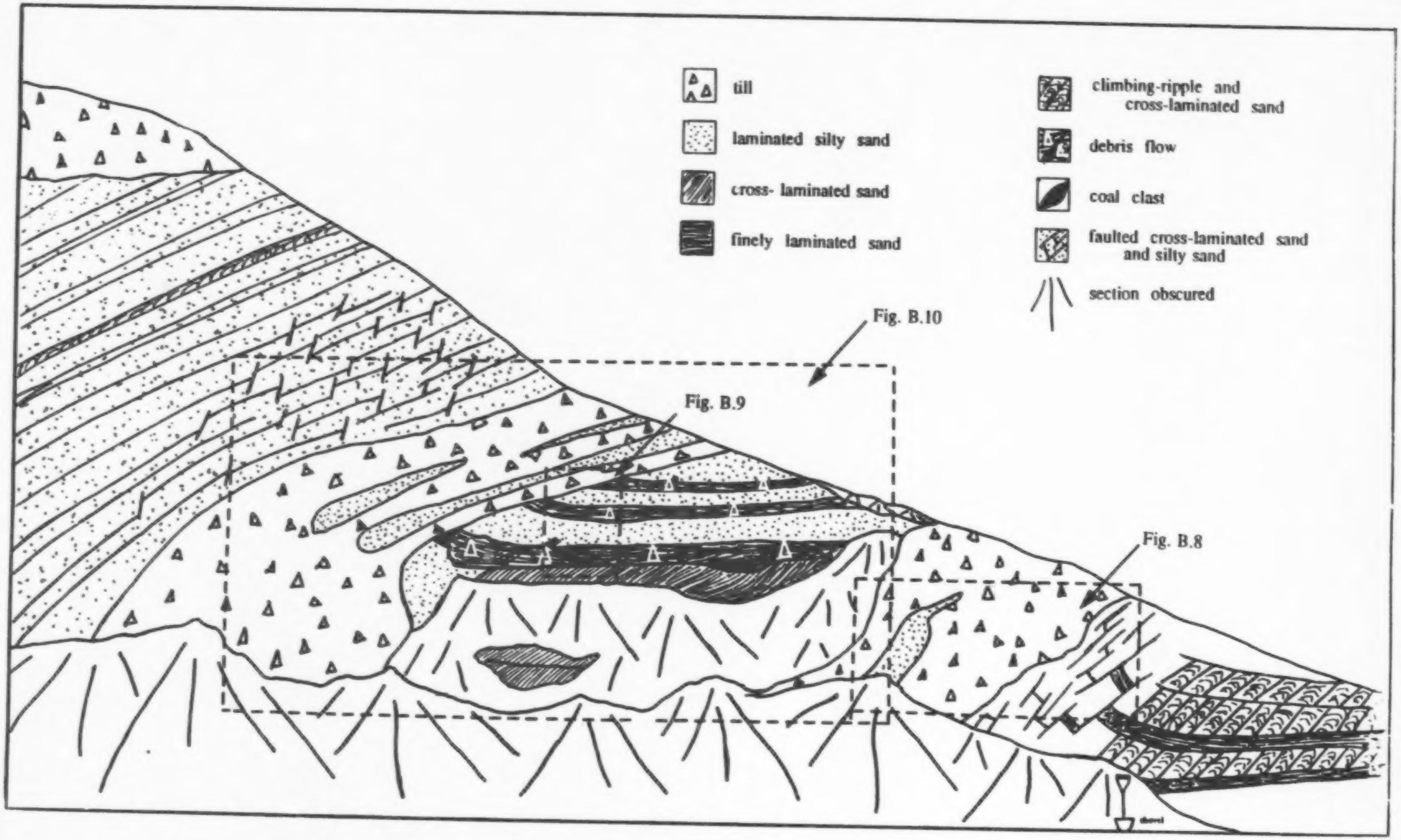

till
laminated silty sand
cross- laminated sand
finely laminated sand
climbing-ripple and cross-laminated sand
debris flow
coal clast
faulted cross-laminated sand and silty sand
section obscured
Fig. B.10
Fig. B.9
Fig. B.8
shovel

Figure B.7: A schematic diagram of section 4 on Romulus lowlands. See text for discussion. Shovel is *ca.* 1 m long.

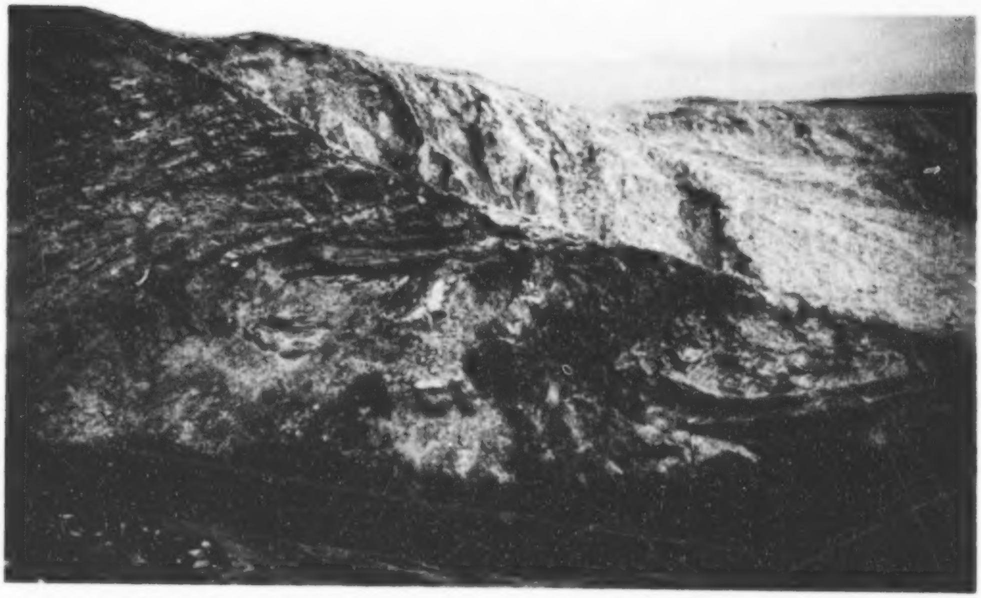


Figure B.7: Section 4 in tributary valley of Romulus river. The upper till does not appear in this photo. The lowermost sand unit is located behind the shovel (1 m long) in the right foreground.

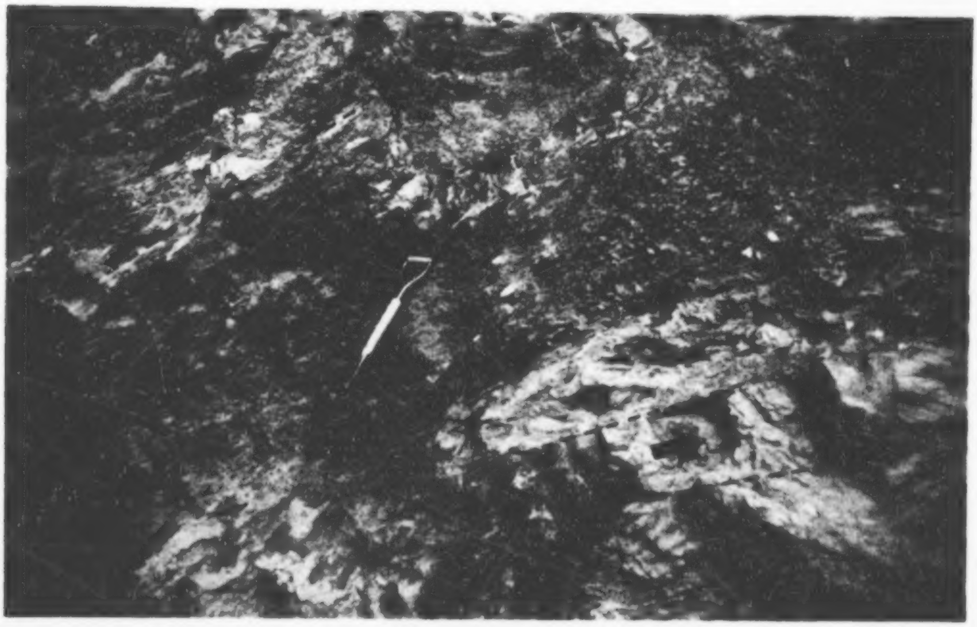


Figure B.8: View of the lowermost sand and the overlying till. Note the displacement of sand upslope along thrust faults (arrow shows direction) and the intratill sand raft (dashed line above shovel).

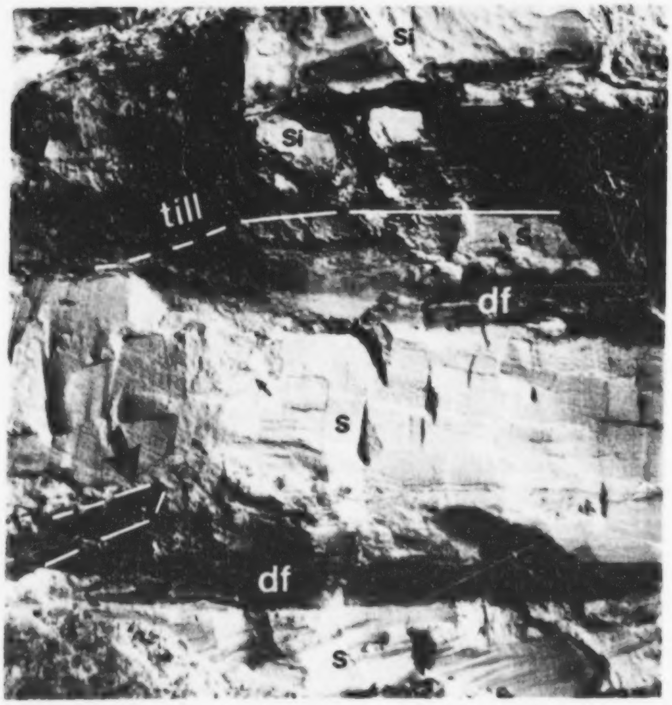


Figure B.9: View of interbedded debris flows (df) and channel/tunnel sand (s) truncated by overlying till (dashed line) with sand intrabed (si). The lowest debris flow has been thrust into the overlying sand (large arrow). Note silt injection along faults in sand (small arrows).

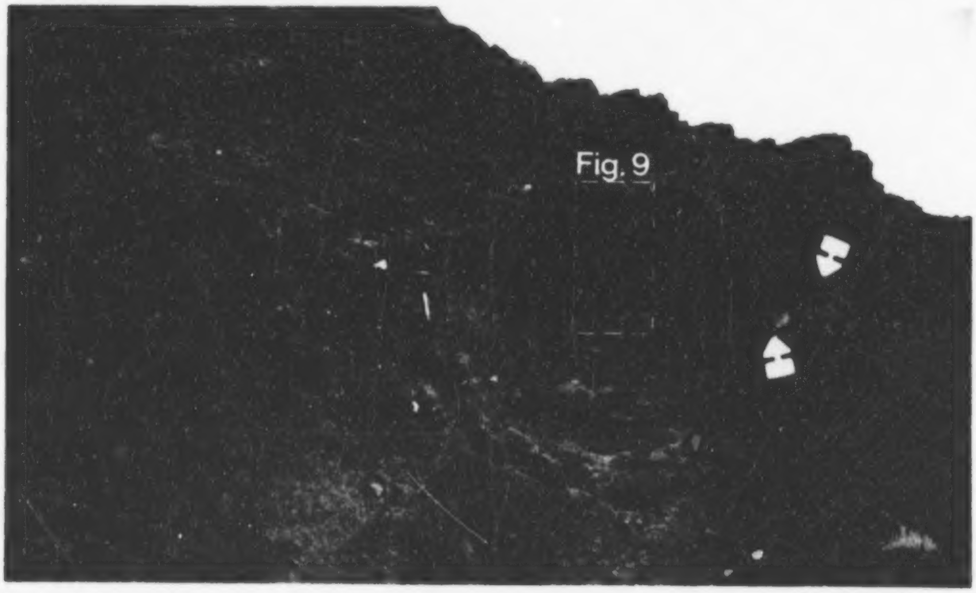


Figure B.10: View of the central part of section 4 showing interbedded debris flows and channel/tunnel sand (right), second till (left), and upper sand (top). Note boulder and coal clast (arrows).

Overlying the till is 6-12 m of cross-laminated sand and sandy silt containing lenses of stony diamicton, coal and unaltered wood (Figs. B.7 and B.10). The sand conforms to the upper contact of the till and is pervasively (normal) faulted immediately overlying the contact. This unit represents the progradation of fan or deltaic deposits over the site following ice retreat. The inclination of the sand beds suggest a sediment source to the east or northeast. The presence of coal clasts and unaltered wood is consistent with this source as coal outcrops in Eureka Sound Group bedrock east of the site; late Tertiary wood is preserved in high terrace sediments to the east and southeast (Figs. 2.6 and 4.1). Pervasive normal faulting of the sand is attributed to subsidence and thaw-consolidation during melt-out of the underlying till. Lenses and thin units of diamicton within the sand may be related to either localized debris flows on the delta front or ice-rafted debris.

Capping the section is a partially indurated, stony diamicton which unconformably overlies the bedded sand (Fig. B.11). The abrupt, planar diamicton-sand contact overlies a zone of penetratively deformed sediments (Fig. B.12). The original sedimentary structure has generally been replaced by a heterogenous bed of deformed sand and silt. Below this zone, laminated sand has been folded and faulted with silt injected along fracture planes. The base of the diamicton contains intrabeds of laminated brown sand, sometimes outlined by black coal grains. These thin laminae have been stretched, forming boudins and are normally faulted (Fig. B.12). The diamicton is interpreted as till. It is laterally continuous across the site and occurs as a blanket on the adjacent lowlands. Subtill deformation in the sand is consistent with ice loading.

The stratigraphy in section 4 indicates an advance of ice into a subaqueous fan or deltaic environment. The lowest sand unit and the thick inclined sand beds represent a distal to proximal transition in a fan/delta prograding westward onto the lowland. Ice-directional indicators in the two lowest tills suggest an ice advance from Eureka Sound although neither appears to be fossiliferous. The tills and interbedded debris flows and channel/tunnel sand represent either a fluctuating ice front with ice-marginal sedimentation or subglacial deposition during a single ice advance. The uppermost till outcrops discontinuously on the lowland and is fossiliferous nearer the Eureka Sound coast. It represents the most recent glaciation of this area.

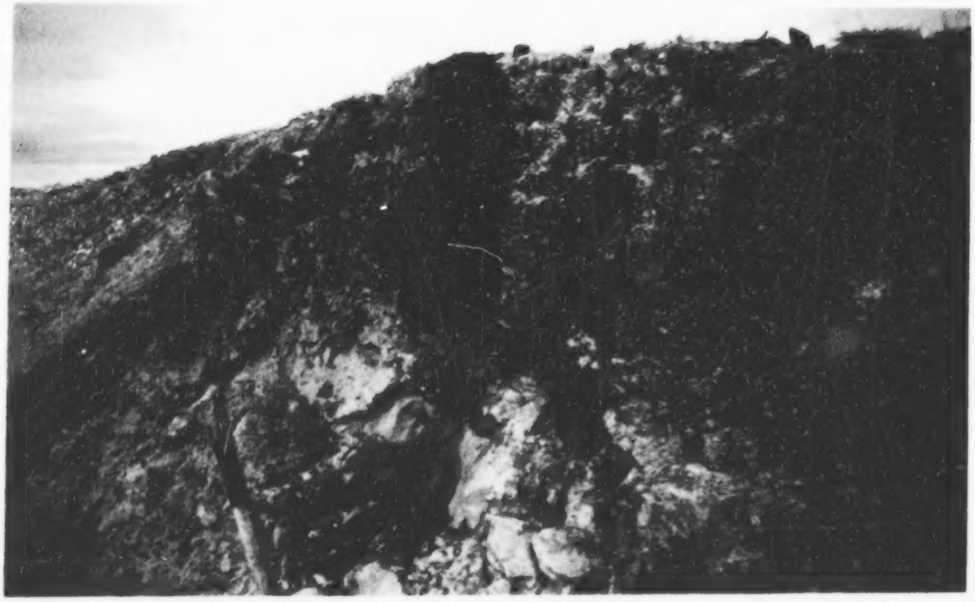


Figure B.11: View of the upper till that caps section 4. This till occurs as a discontinuous veneer on Romulus lowland.

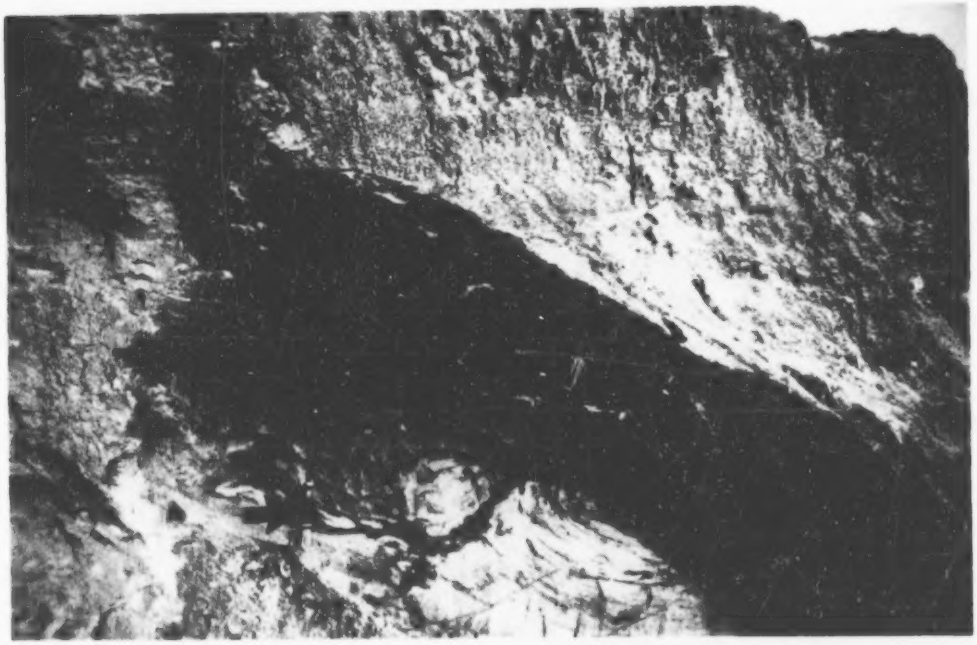


Figure B.12: View of the penetratively deformed zone below the upper till in section 4. Above the knife (arrow), deformation has resulted in a heterogeneous sand and silt unit whereas weaker deformation below the knife has caused shearing, silt injection and folding.

SECTION 5, HOT WEATHER CREEK

Near Hot Weather Creek, a retrogressive thaw slump (Fig. B.13) exposed frozen diamicton in its 3 m high headwall (Fig. B.14). The diamicton is composed of a silty clay matrix with subrounded striated clasts (predominantly quartzite and carbonate). The most striking feature of the diamicton is the incorporated sand rafts (Fig. B.15). The largest raft, measuring 1 m by 2 m, appears to have been rotated and streamlined in a westerly direction. The sand is folded and faulted but generally retains its primary structure. Clasts and small stringers of sand in the diamicton are oriented parallel to raft edges (Fig. B.15).

Towards the west (left, Fig. B.13), stacked rafts of alternating grey and brown sand (30-50 cm thick) are inclined at 20° within the diamicton (Fig. B.16). Thin diamicton beds (3-10 cm thick) outcrop between the sand rafts, and pebbles are oriented parallel to the inclined beds. A thin sand lamination emanates from the base of one of these sand rafts and can be traced into the overlying diamicton where it becomes subhorizontal. Tabular and trough cross bedded sand was exposed in a gully below the retrogressive thaw slump (Fig. B.13) and similar deposits occur in isolated sections in lower Hot Weather Creek valley. The sand is weakly deformed (mostly faulted) and appears similar to the rafts in the diamicton.

A thin sand bed (2-5 cm) drapes the upper contact of the diamicton and is overlain by a cryoturbated silty unit of variable thickness (cm's; Fig. B.14). Shells were observed on the silt surface and a nearby *in situ* sample provided an early Holocene date (site 23).

The diamicton is interpreted to be till deformed by subglacial shear. Rafts of sub-till sand have been incorporated into the till and because they are more competent than the clayey-silt till matrix, they have been rotated in the direction of shear (westward). Subglacial shearing has also caused stacking of sand rafts with attenuation of sand laminae in a down-ice direction. The overlying sand and fossiliferous silt appear to drape the till surface, much like an overstep sequence (Stewart 1991); however, the thin sand unit may represent an erosional hiatus in the sedimentary record, suggesting two discrete events. The orientation of subglacial shearing suggests ice movement from the east or southeast.

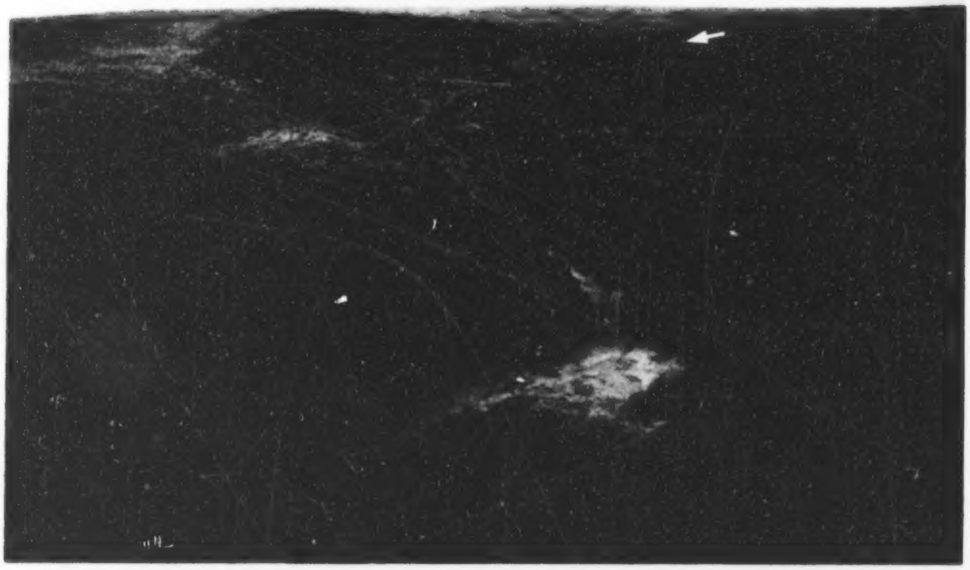


Figure B.13: Section 5 exposed in the headwall of a retrogressive thaw slump near Hot Weather Creek. Blue backpack (arrow) for scale. Note sand exposed in gully at bottom left.

Figure B.14: View of the right side of section 5 showing sand rafts in till martrix. The largest raft measures 1 by 2 m. The thin sand unit that drapes the till can be seen at top left (arrow).

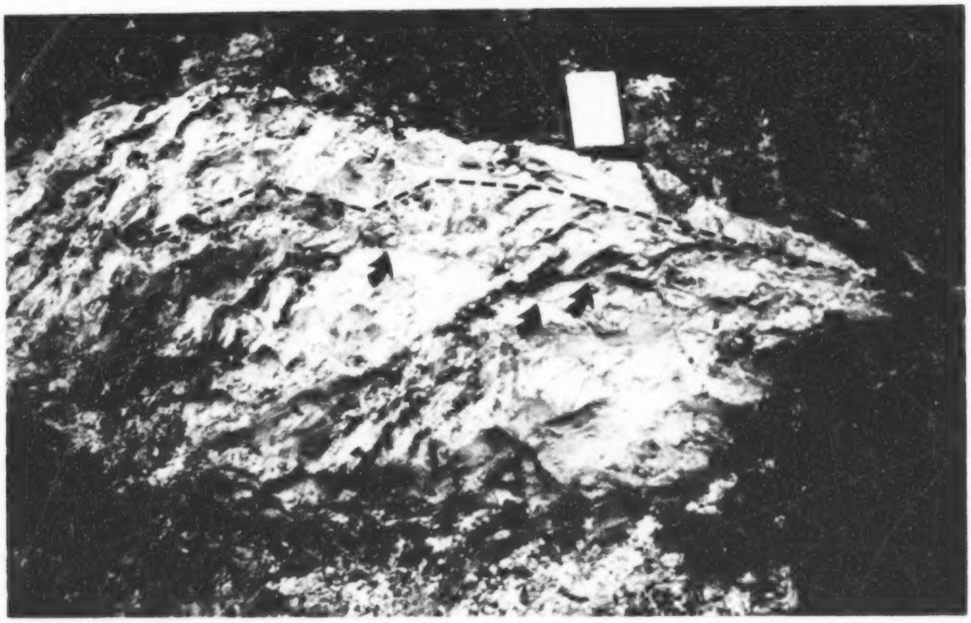


Figure B.15: View of large sand raft in section 5. Although the sand has been folded and faulted (arrows), it retains its primary structure. Sand from the nose of the raft (left) has been smeared over its upper surface (dashed line). Field notebook for scale.

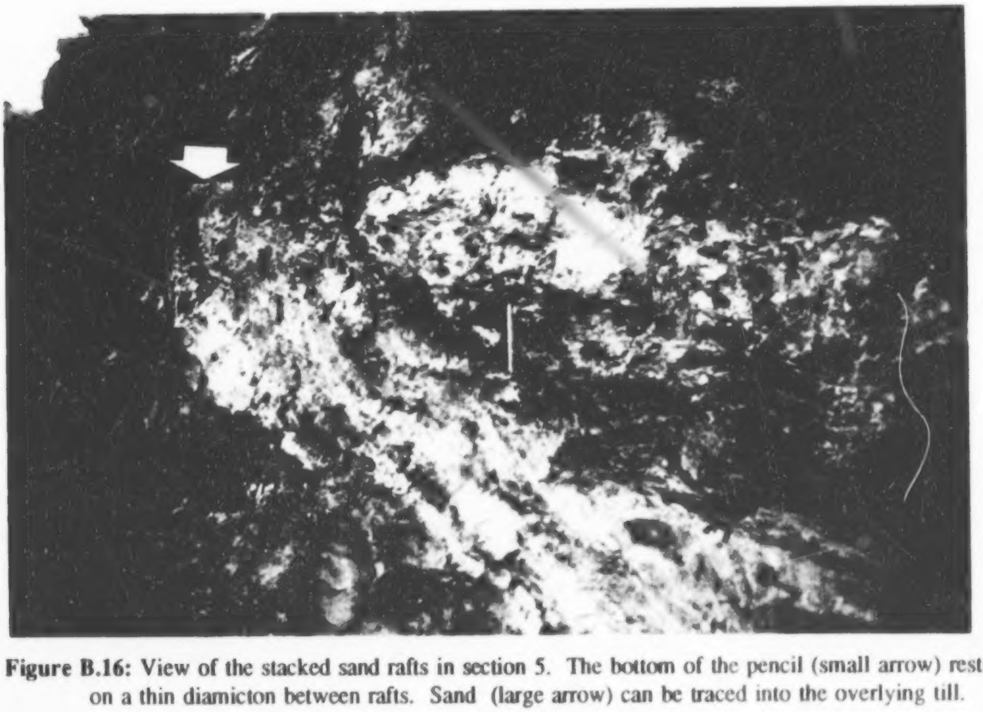


Figure B.16: View of the stacked sand rafts in section 5. The bottom of the pencil (small arrow) rests on a thin diamicton between rafts. Sand (large arrow) can be traced into the overlying till.

APPENDIX C

Radiocarbon dates and amino acid ratios

Table C.1: Radiocarbon dates obtained for this study

Site No.[a]	Lab. No.[b]	Material[c]	Weight used (g)[d]	Age[e]	Preservation and enclosing material[f]	Sample elev. (m)[g]	Related sea level (m)[h]	Lat. N	Long. W	Geological environment[i]
Troelsen ridge										
1	TO-2233	Shells, *M.t.*	.347	8 440 ± 80	valves; silt surface	94	>94 ≤152	80°11′	86°33′	Nearshore deposits
2	TO-2235	Shells; *H.a.*	.230	8 880 ± 90	valves; silt surface	110	>110 ≤152	80°04′	86°34′	Nearshore deposits
3	TO-2231	Shells, *M.t.*	.286	8 210 ± 80	*in situ*; silt	59	>59 ≤152	80°05′	86°33′	Prodeltaic deposits
Hare Cape ridge										
4	GSC-5119	Shells; *H.a.*	33.8	7 920 ± 100	valves; stony silt	75	>75 ≤152	79°51′	86°25′	Glaciomarine deposits
5	GSC-5156	Shells; *H.a.*	32.4	8 680 ± 80	*in situ*; stony silt surface	132	>132 ≤152	79°48′	86°18′	Nearshore deposits
5	TO-2237	Shells; *H.a.*	.204	8 640 ± 80	*in situ*; stony silt surface	132	>132 ≤152	79°48′	86°18′	Nearshore deposits
6	TO-2238	Shells; *M.t.*	.270	8 280 ± 80	*in situ*; silt surface	89	>89 ≤149	79°46′	85°52′	Prodeltaic deposits
7	TO-1242	Shells; *M.t.*	.329	8 230 ± 70	*in situ*; silt	99	>99 ≤149	79°44′	85°47′	Prodeltaic deposits
Romulus lowland										
8	TO-2241	Shells; *M.t.*	.286	8 480 ± 80	*in situ*; clayey silt	93	>93 ≤147	79°50′	85°13′	Nearshore deposits
9	GSC-4957	Shells; *M.t.*	41.4	8 230 ± 80	valves; stony silt surface	110	>110 ≤147	79°51′	85°17′	Nearshore deposits
Black Top Ridge / Snow Goose hill										
10	GSC-4708	Shells; *M.t.*	36.8	8 520 ± 80	*in situ*; silt	100	>100 ≤147	79°58′	85°22′	Nearshore deposits
11	TO-2242	*Salix* twig	.032	44 420 ± 840	Detrital; sand	34	-	80°15′	84°39′	Glaciomarine deposits
11	TO-2236	Shells; *P.a.*	.285	9 080 ± 80	*in situ*; silt	63	>63 ≤149	80°15′	84°39′	Glaciomarine deposits
11	TO-2243	Shells; *M.t.*	.309	7 890 ± 80	*in situ*; silt	70	>70 ≤149	80°15′	84°39′	Glaciomarine deposits

12	TO-2229	Shells; *M.t.*	.639	8 450 ± 80	*in situ*; stony silt surface	94	>94 ≤149	80°15′	85°02′	Nearshore deposits
Eastwind Lake lowland										
13	TO-2232	Shells; *H.a.*	.208	8 990 ± 80	valves; silt surface	137	>137 ≤149	80°11′	85°28′	Nearshore deposits
14	TO-2230	Shells; *Ast.*	.304	6 670 ± 70	*in situ*; silt surface	60	>60 <149	80°14′	85°18′	Nearshore deposits
Central Eureka Sound										
15	TO-2245	Shells; *H.a.*	.339	8 430 ± 70	valves; silt surface	101	>101	79°14′	85°30′	Nearshore deposits
16	TO-2244	Shells; *M.t.*	.228	6 480 ± 70	*in situ*; silty sand	17	≥30	79°19′	85°32′	Prodeltaic deposits
17	GSC-5147	Shells: *H.a.*	30.6	5 180 ± 90	*in situ*; sandy gravel	14	≥16	79°19′	85°31′	Delta foresets
18	GSC-5117	Shells; *H.a.*	31.6	6 330 ± 90	*in situ*; sandy gravel	28	≥30	79°14′	85°38′	Delta foresets
Central lowlands										
19	TO-2240	Shells; *P.a.*	.244	9 560 ± 90	*in situ*; clayey silt	88	>88 ≤145	79°56′	84°17′	Glaciomarine (Ia) deposits
19	TO-2611	Echinoderm spicules	.035	8 900 ± 90	*in situ*; silty clay	83	>83 ≤145	79°56′	84°17′	Glaciomarine (Ia) deposits
20	TO-1963	Shells; *P.a.*	.435	9 230 ± 100	*in situ*; silty clay	126	>126 ≤145	79°59′	84°05′	Glaciomarine deposits
21	TO-2234	Shells; *M.t.*	.260	8 270 ± 80	*in situ*; fine sand	105	>105 ≤145	79°56′	84°12′	Glaciomarine (Ib) deposits
21	TO-1964	Shells; *M.t.*	.493	8 230 ± 80	paired, silt surface	95	>95 ≤145	79°56′	84°12′	Glaciomarine (Ia) deposits
22	TO-1965	Shells; *P.a.*	.037	5 870 ± 80	*in situ*; silty clay	21	>21 <145	79°53′	84°37′	Glaciomarine (Ia) deposits
23	GSC-5155	Shells; *M.t.*	32.7	8 570 ± 120	*in situ*; silt surface	100	>100 ≤145	79°58′	84°26′	Nearshore deposits
Iceberg Point lowland										
24	TO-2228	Shells	.125	45 850 ± 980	fragmented; sand	98	na	80°15′	86°19′	Glaciomarine (?) deposits
Greely Fiord										
25	TO-1966	Narwhal tusk	2.5	7 970 ± 90	fragment; diamicton	5	>5 ≤149	80°17′	84°18′	Colluvium

a Numbers correspond to locations in Fig. 4.10.

b The laboratory prefixes TO and GSC designate samples dated by AMS at IsoTrace, University of Toronto and by conventional analysis at the Geological Survey of Canada, respectively.

c Shells: *H.a.*, *Hiatella arctica*; *M.t.*, *Mya truncata*; *Ast.*, *Astarte* sp.; *P.a.*, *Portlandia arctica portlandia* Hitchcock, identified by A. Aitken, and considered by Lubinsky (1980) to be characteristic of shallow, less saline epicontinental seas.

d Shell samples received a 30-50% preleach except GSC-4957, 5117, 5119, 5147, 5156 and TO-1963 to 1966 which had only the outer 20% removed.

e GSC shell dates have been corrected for fractionation to a base of $\delta^{13}C = 0‰$, except GSC-5117, 5119 and 5147 where the uncorrected (machine) age is given. All IsoTrace dates have been corrected for fractionation to a base of $\delta^{13}C = -25‰$; however, these shell dates have also been adjusted for a reservoir effect of 410 years which is roughly equivalent to a fractionation correction to a base of $\delta^{13}C = 0‰$. IsoTrace and GSC date errors represent one and two standard deviations, respectively.

f Shell samples described as *in situ* were found as paired bivalves in foreset beds or in suspension-load sediments.

g Altimeter readings have been corrected for temperature and pressure changes.

h As the contemporary sea level is not known for any of the dated samples, it is assumed to lie above the elevation of each sample site but at or below the local Holocene marine limit. Note: na = not applicable.

i Bold letters denote lithofacies type. See text for description.

Table C.2: Amino acid ratios obtained for this study

Site No.[a]	Lab No.[b]	Genera[c]	alle/Ile ratios[d]		Thermal[e] category	Enclosing material	Sample elev. (m)	Lat. N	Long. W	Comments
			Free	Total						
						GROUP I				
Iceberg Point lowland										
26	AGL-1688A	*H.a.*	0.593	0.133	3	Till surface	198	80°13′	86°17′	Abundant granite erratics on till surface.
	B	*H.a.*	0.620	0.106						
	C	*H.a.*	0.466	0.107						
27	AGL-1704A	*M.t.**	0.555	0.185	1	Till	134	80°14′	85°57′	Granite erratics present. See Fig. 3.18 for photograph of exposure.
	B	*M.t.**	*0.510*	*0.058*						
	C	*M.t.**	0.712	*0.300*						
	D	*H.a.**	0.606	0.173						
28	AGL-1710A	*H.a.*	0.465	0.091	1	Till (?)	98	80°15′	86°15′	Till (?) overlies pebbly sand (AGL-1862). See section #3. Granite erratics present.
	B	*M.t.*	0.372	0.093						
	C	*Ast.**	*0.532*	*0.130*						
28	AGL-1862A	frags.	0.599		1	Pebbly sand	98	80°15′	80°19′	45 850 ± 980 (TO-2228)
Eastwind Lake lowlands										
29	AGL-1684A	*H.a.*	0.408	0.089	2	Till surface	157	80°10′	85°30′	Sample included whole valves. No granites observed.
	B	*H.a.*	0.480	0.109						
	C	*H.a.*	0.482	0.096						
30	AGL-1863A	*M.t.*	0.741	0.164	2	Till surface	165	80°11′	85°00′	Till overlain by deltaic sand and gravel related to former sea level at 170 m asl.
	B	*M.t.*	0.650	0.182						
	C	*Ast.*	*0.646*	*0.215*						
Greely Fiord										
31	AGL-1700A	*M.t.**	0.400	0.118	1	Beach sand and gravel	129	80°13′	84°50′	Beach material reworked from underlying till. No granites observed.
Central Eureka Sound										
32	AGL-1701A	*H.a.*	0.501	0.118	3	Till surface	225	79°19′	85°32′	No granite observed.
	B	*H.a.*	0.564	0.144						
	C	*H.a.*	0.604	0.116						
	D	*H.a.*	0.588	0.130						

GROUP II

Hare Cape ridge										
33	AGL-1707A	*M.t.*	0.206	0.065	3	Till surface	612 - 626	79°54′	86°20′	30 000 ± 1600 BP (GSC-111) 36 300 ± 2000 BP (GSC-111) 28 700 ± 600 BP (GSC-51)
	B	*M.t.*	0.163	0.064						
	C	*M.t.*	0.204	0.068						
34	AGL-1687A	*M.t.**	0.242	0.047	3	Till surface	324	79°54′	86°13′	Granite erratics observed
	B	*M.t.**	0.146	0.079						
	C	*M.t.**	0.237	0.059						
35	AGL-1685A	*M.t.*	0.194	0.064	3	Till surface	272	79°51′	86°00′	19 500 ± 1100 BP (L-548) Sim (1961) Granite erratics observed.
	B	*M.t.*	0.209	0.065						
	C	*M.t.*	0.194	0.069						
36	AGL-1690A	*H.a.*	0.241	0.076	1	Till surface	145	79°45′	85°12′	Granite erratics observed.
	B	*H.a.*	0.265	0.050						
	C	*H.a.*	0.234	0.086						
	D	*M.t.*	0.238	0.064						
	E	*M.t.*	0.355	0.083						
Fulmar Channel										
37	AGL-1686A	*H.a.*	0.262	0.064	3	Till surface	469	79°11′	86°02′	Granite erratics present.
	B	*H.a.*	*0.875*	*0.048*						
	C	*M.t.*	0.277	0.065						
	D	*M.t.*	0.284	0.045						
Romulus Lowland										
38	AGL-1689A	*M.t.*		0.059	1	Till surface	96 - 116	79°46′	85°39′	Valves more fragmented inland from Eureka Sound. Rare granites present.
	B	*M.t.*		0.067						
	C	*M.t.*		0.052						

GROUP III

Greely Fiord										
39	AGL-1706A	*H.a.*	0.285	0.049	1	Till surface	142	80°14′	84°16′	No granites observed.
	B	*H.a.*	0.379	0.041						
	C	*H.a.*	0.358	0.089						
Central Eureka Sound										
40	AGL-1703A	*Ast.*	*0.200*	*0.059*	1	Till surface	174	79°17′	85°37′	No granites observed. Till overridden by last ice advance.
	B	*Ast.*	*0.615*	*0.055*						
	C	*Mya/H.a.*	0.417	0.090						

INTERNATIONAL LABORATORY STANDARDS (March-June 1992)

	Free	Total
ILC-A Saxidomus	0.439	0.161 ±.0056 (n=2)
ILC-B Mercenaria	0.940	0.500 ±.0300 (n=2)
ILC-C Mercenaria	1.320	1.052 ±.0250 (n=2)

AGL ANALYTICAL STANDARDS

For the periods:		
3/24 - 4/16, 1992	0.097 ±.0020	C.V. = 2.1%
4/23 - 6/6, 1992	0.234 ±.0045	C.V. = 1.8%

---

a Numbers correspond to locations in Figure 4.10.

b Shell samples were analyzed at the Amino Acid Geochronology Laboratory, University of Massachusetts. The letters A-E refer to individual valves or fragments analyzed from each sample.

c Shells: *H.a.*, *Hiatella arctica*; *M.t.*, *Mya truncata*; *Ast.*, *Astarte* sp.. An asterisk denotes fragments that were identified by their characteristic amino acid content.

d Each ratio is a mean value for at least three individual analysis on each valve or fragment. The ratios in italics were not used in data reduction or analysis because of (1) taxonomic differences, (2) anomalously high ratios from 'old' shells, or (3) inconsistent free and total ratios suggesting analytical error.

e Refers to sample elevation in relation to former marine transgressions: 1 = below Holocene marine limit (≈150 m asl); 2 = above limit of Holocene sea but below marine limit (170-185 m asl); 3 = above marine limit.

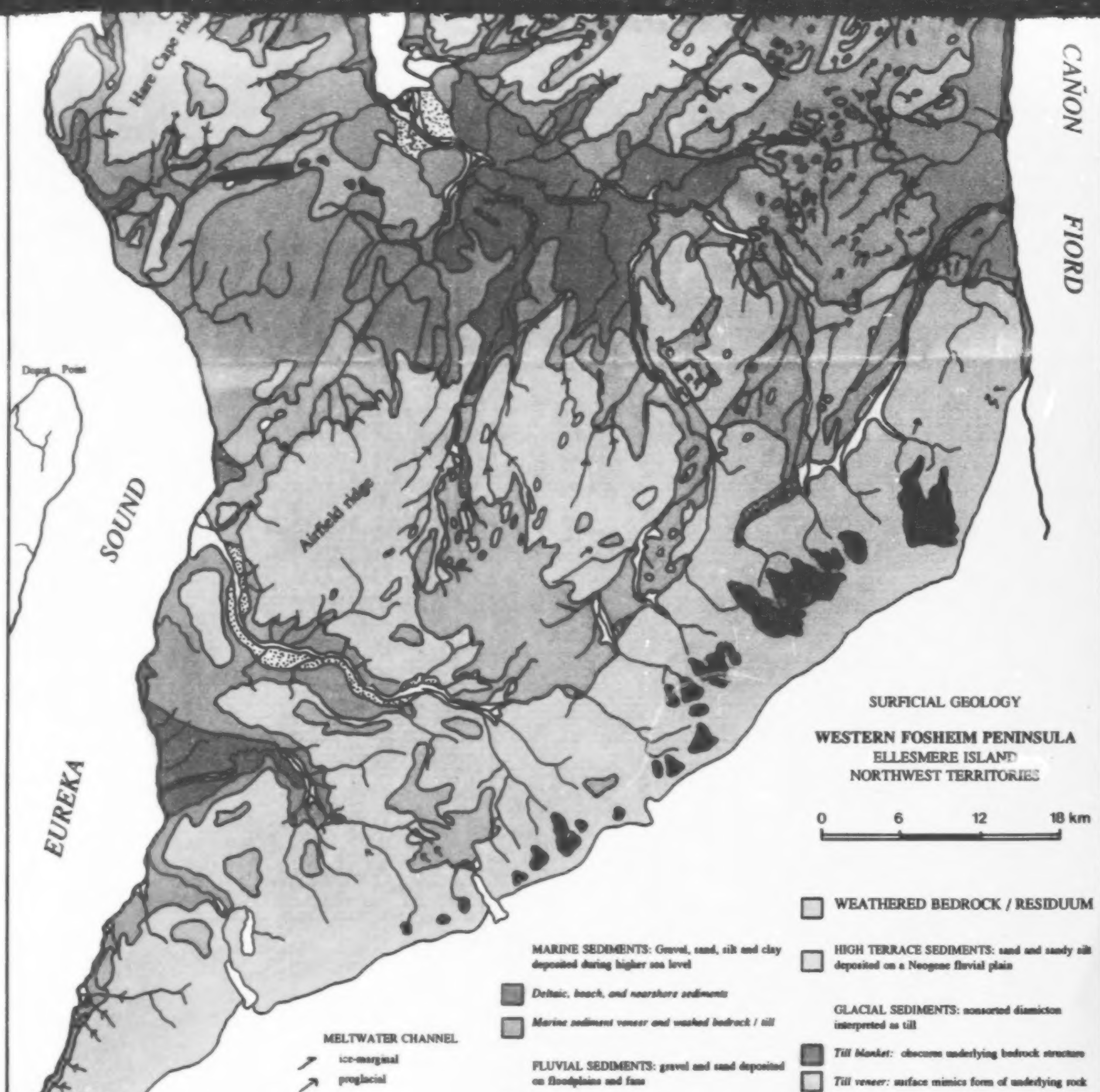
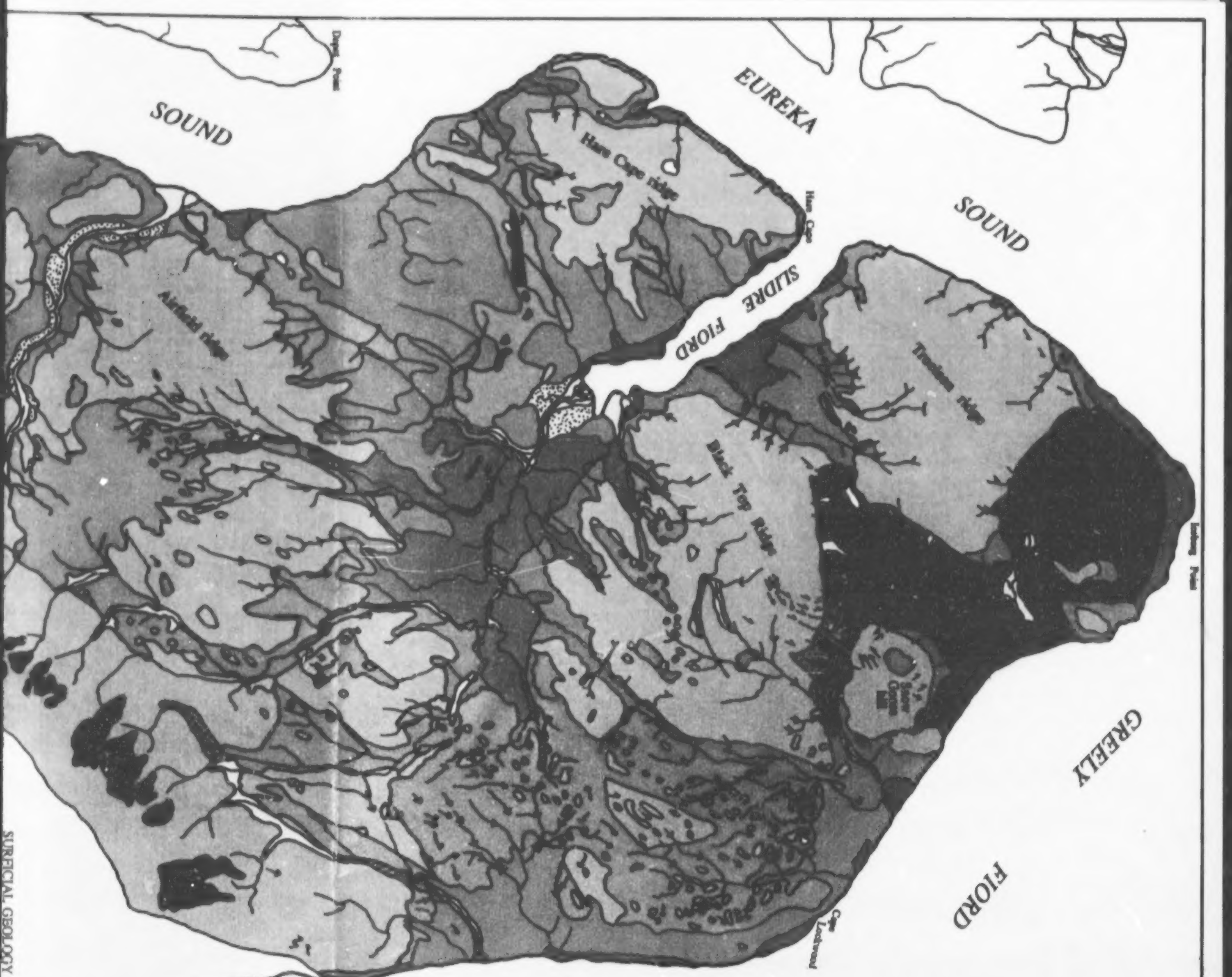

CAÑON
FIORD
Hare Cape ridge
Depot Point
SOUND
Airfield ridge
EUREKA
VESLE FIORD
SURFICIAL GEOLOGY
WESTERN FOSHEIM PENINSULA
ELLESMERE ISLAND
NORTHWEST TERRITORIES
0
6
12
18 km
WEATHERED BEDROCK / RESIDUUM
HIGH TERRACE SEDIMENTS: sand and sandy silt deposited on a Neogene fluvial plain
GLACIAL SEDIMENTS: nonsorted diamicton interpreted as till
Till blanket: obscures underlying bedrock structure
Till veneer: surface mimics form of underlying rock
GLACIOFLUVIAL SEDIMENTS: gravel and sand deposited in the marginal zone of a former glacier
Proglacial outwash
MARINE SEDIMENTS: Gravel, sand, silt and clay deposited during higher sea level
Deltaic, beach, and nearshore sediments
Marine sediment veneer and washed bedrock / till
FLUVIAL SEDIMENTS: gravel and sand deposited on floodplains and fans
Seasonally flooded sediments
Terraced sediments above present floodplain
GLACIER ICE
MELTWATER CHANNEL
ice-marginal
proglacial
MORAINE
DELTA

Figure 3.1: Surficial geology and geomorphology of western Fosheim Peninsula.


SOUND
Depot Point
EUREKA
Hare Cape ridge
Hare Cape
SOUND
SLIDRE
FIORD
Airfield ridge
Troelsen ridge
Black Top Ridge
Iceberg Point
GREELY
FIORD
Cape Lockwood
CAÑON
FIORD
SURFICIAL GEOLOGY

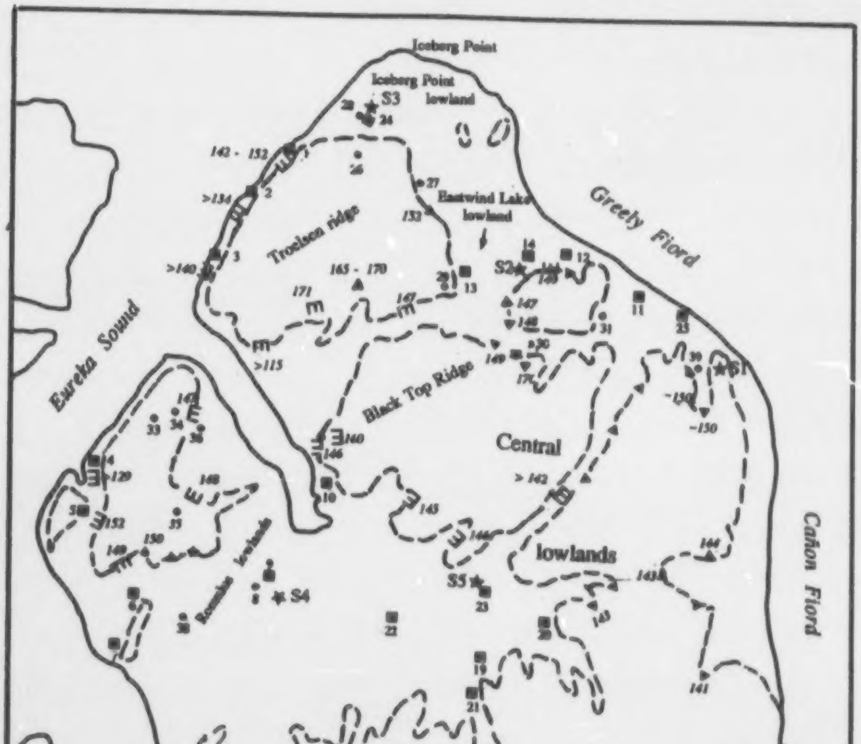

Iceberg Point
Iceberg Point lowland
S3
23
24
26
27
142 - 152
1
>134
2
3
>140
Troelsen ridge
165 - 170
171
147
132
29
13
Eastwind Lake lowland
Greely Fiord
14
12
S2
144
125
147
148
31
11
25
>115
149
30
17
39
S1
-150
-150
Eureka Sound
Black Top Ridge
Central
lowlands
142
33
34
36
140
146
10
4
120
148
S
152
35
150
148
Romulus lowlands
145
>142
144
S5
23
143
144
S4
9
38
22
20
143
19
21
141
-140
164
Central Eureka Sound
32
17
16
40
15
>101
18
0
18
km
^{14}C shell sample site
amino acid shell sample site
S1
stratigraphic section # 1
delta (elevation m)
highest marine feature (elevation m)
Holocene marine limit (approximate)

Figure 3.10: Location of shell sample sites, stratigraphic sections, raised marine features, and marine limit (approximate).